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THE LATE PLEISTOCENE SUBMERGENCE IN THE COLUMBIA VALLEY OF OREGON AND WASHINGTON

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This article aims to review previous references to a Champlain submergence in the lower drainage of Columbia River, to eliminate erroneous interpretations and correlate data which belong to this subject, to present more evidence establishing the fact, to define the character and the extent of the submergence, to correlate this with the Pleistocene history of northwestern Washington, and to indicate questions which are still unanswered.

HISTORICAL

So far as known, Thomas Condon published the first note on static water in the Columbia Valley during late Pleistocene time. This was an article entitled "The Willamette Sound" in the *Overland Monthly*, 1871.[1] Condon argued for a submergence of at least 330 feet above present tide, the waters flooding Willamette Valley and backing up through the Gorge of the Columbia across the Cascade Range and submerging portions of the tributary Yakima and Walla Walla valleys. Condon's evidence was the

[1] The article has been reprinted as a chapter in Condon's *The Two Islands* (Portland: J. K. Gill, 1902), and also in his *Oregon Geology* (1910), a revision of *The Two Islands*.

existence of marine Pleistocene deposits along Shoalwater Bay, Washington, 10 to 15 miles north of the mouth of Columbia River, and fossiliferous beds below about 330 feet A.T. at the mouth of Des Chutes River, an Oregon tributary entering the Columbia immediately east of the Cascade Range. If the sediments in these two localities, more than 150 miles distant from each other, were deposited in the same body of water, then Willamette Valley of western Oregon, intermediate in position, must have been largely submerged. To this water body Condon gave the name of Willamette Sound. He correlated it with the Champlain submergence on the Atlantic Coast.

Static waters in the Columbia Valley east of the Cascade Range were postulated by T. W. Symons in a Senate document,[1] published in 1882, dealing chiefly with navigation of the Columbia. Symons saw many rounded bowlders on the basalt plateau along the Columbia in central Washington, he saw considerable tracts of the basalt covered with sedimentary deposits, and he found large, though fragmentary, gravel terraces in the Columbia Valley. All of these he ascribed to a glacial lake which he named Lake Lewis. He thought this lake extended southward as far as Walla Walla and Wallula. He gave no altitudes, cited no evidence for the Champlain age which he assumed for the submergence, and suggested no cause for the ponding.

I. C. Russell made a geological reconnaissance in central Washington in 1892, and in his report[2] revised Symons' data somewhat but adopted his conception of a glacial lake. His revision consisted in eliminating the sedimentary formation exposed in White Bluffs along the Columbia. This he considered to be a portion of the John Day Series of Tertiary age. He also added the valuable item that there were many foreign granite bowlders in the Columbia and Yakima valleys which had been carried in bergs to their present positions. Russell thought that a glacier might have dammed Columbia River at The Dalles, just below the mouth of Des Chutes River, but he admitted the possibility of subsidence

[1] "The Upper Columbia River and the Great Plain of the Columbia," *Senate Document No. 186*, Forty-seventh Congress, 1st sess., Washington, 1882.

[2] "A Geological Reconnaissance in Central Washington," *U.S. Geol. Surv., Bull. 108* (1893).

allowing "the ocean to enter the great central valley between the Cascades and Rocky mountains."

J. S. Diller published his geological reconnaissance of northwestern Oregon[1] two years after the appearance of Russell's report. In it he adopted Condon's notion of a Willamette Sound. Like Russell, he announced the existence of foreign bowlders in the valley and interpreted these as berg-carried in the waters of a Pleistocene submergence.

Russell, in another paper, published in 1898,[2] stated his belief that the Great Terrace in the Columbia Valley, between the junction of the Methow and Chelan valleys, was a delta built in Lake Lewis. He considered that the termination of the Great Terrace, a little below the mouth of Chelan River, was the delta front. Russell noted here that "full and unquestionable evidence of its [the lake's] actual existence cannot be said to have been discovered."

CRITICAL EXAMINATION OF THE DATA PRESENTED BY THE FOREGOING WRITERS

Shoalwater Bay.—These deposits, cited by Condon, are unquestionably part of the coastal phase of the Satsop formation.[3] This formation is largely a river gravel in the Columbia Valley, and only along the coast was it deposited in standing water. In the Shoalwater Bay (Willapa Bay) exposures there is abundant evidence in the presence of gravel strata containing granite, gneiss, and much quartzite that a vigorous current existed throughout the length of the Columbia in Oregon and Washington when these beds were deposited. Though the Satsop is believed from paleontological evidence to be Quaternary in age, it was laid down before the present Cascade Range was folded and therefore long antedates the Champlain epoch. The strata along Shoalwater Bay therefore do not record a Willamette Sound.

[1] "A Geological Reconnaissance in Northwestern Oregon," *U.S. Geol. Surv., 17th Ann. Rept.*, Part I (1895).

[2] "The Great Terrace of the Columbia," *Amer. Geologist*, XXII (1898), 363.

[3] J H. Bretz, "The Satsop Formation of Oregon and Washington," *Jour. Geol.*, XXV (1917), 447.

Des Chutes terrace.—Condon gave no list of fossils from the terrace deposit. He simply said the material was stratified clays, sands, and gravels containing tusks, teeth, and bones "of the land animals of that period, marking at once the height at which these waters stood." No one has since studied the fauna from this locality. There are many Tertiary formations, with vertebrate faunas, in Oregon and Washington in the great lava plain east of the Cascades. The terrace may be the outcropping edge of a formation of this kind, and not a remnant of a post-valley filling. The feature, so far as known, gives no positive evidence for the existence of Willamette Sound.

The Dalles, Oregon.—Condon mentioned sediments deposited in ravines cut in The Dalles beds at this place and ascribed them to Willamette Sound.

Swan Island, Portland.—Quoting Condon: "Currents of such a vast body of water [Willamette Sound] . . . . the agency competent to the heaping up of that long sandy ridge, 100 feet high, through which the river has cut at Swan Island, north of Portland." This ridge is a terrace remnant; left by erosion, not deposition; and is a portion of a great delta of the Columbia built in Willamette Valley. Static waters stood no higher than its surface (300 feet A.T.) when the delta was built. If related to Willamette Sound it is probably a record of the subsiding stages. It will be discussed more fully later in this paper.

White Bluffs (Ringgold formation), Washington.—Merriam and Buwalda recently have shown[1] that the sedimentary formation exposed in White Bluffs is of very late Cenozoic age, probably Pleistocene. They favor the view that it is a flood-plain deposit, and not lacustral. Its summit plane is considerably below the upper limit of drifted erratics, and on this point it cannot be ruled out of the list of phenomena bearing on the Columbia submergence that we are discussing. But it rises 300 feet higher than the broad gravel terraces on the opposite side of the river, and these are very probably deposits of the river immediately following the Champlain submergence. Merriam and Buwalda found no bowlders of any

[1] "Age of Strata Referred to the Ellensburg Formation in the White Bluffs of the Columbia River," *University of California, Dept. Geol. Bull.*, X (1917), 255.

kind in the materials exposed in the 500-foot sections they studied. Absence of anything attributable to floating ice, presence of a fauna of land vertebrates, and the existence of a broad valley cut in the Ringgold formation containing gravel deposited probably during subsiding stages of the submergence seem adequate evidence for ruling the Ringgold formation out of the catalogue of records of Lake Lewis.

"*Sediments north of Saddle Mountains.*"—It is difficult to determine just what Symons meant by this phrase. There are sediments in that region which are warped with the basalt, and there are horizontal sediments, noted as lake beds of Pleistocene age,[1] which lie below the upper limits of this submergence. These are very poorly exposed and best known from well records. There are also fluvioglacial gravels, deposited by the diverted Columbia during subsiding stages of the submergence and correlative with the gravel terraces of the present Columbia Valley.

"*Rounded bowlders and a loose, light, powdery soil.*"—These occur south of Badger Mountain and Crab Creek, surrounding Saddle Mountain, and in the lower portions of the valleys of Yakima River, Walla Walla River, Snake River, and Moses Lake. Symons here grouped a congeries of deposits of varied genesis: fluvial, fluvioglacial, static water, and wind.

Gravel terraces (cited by Symons).—These are fluvial and fluvioglacial deposits and probably only the Great Terrace, above noted, can be correlated with Lake Lewis.

Neither Condon's nor Symons' work was more than hasty reconnaissance. No one then knew that the sediments of many successive Tertiary basins are exposed in the region, and neither observer was carefully discriminative. Neither one noted the existence of *erratics* in the Columbia Valley. Correlation with the Champlain epoch of the eastern states was with both wholly a matter of inference, on the assumption that the Pacific Coast must have been submerged if the Atlantic Coast was.

Russell (1893) was the first to recognize the significance of the erratic bowlders as evidence for the existence and the age of the

[1] A. T. Schwennesen and O. E. Meinzer, "Ground Water in Quincy Valley, Washington," *U.S. Geol. Surv., Water Supply Paper 425-E*, 1918.

ancient water body. He believed that they were derived from the Okanogan Lobe of the Cordilleran ice sheet. He found these stranded erratics from the mouth of Wenatchee Valley to the lower portion of Yakima Valley.

Diller (1895) cited the erratic bowlders of the Willamette Valley as his contribution to the evidences for this Plęistocene submergence. As in Russell's amplification of Symons' idea of a lake, Diller supplied the only trustworthy evidence (and the convincing proof) of the existence of a late Pleistocene water body in the Willamette Valley. Diller, however, thought that the berg-carried bowlders were derived from glaciers descending the western slope of the Cascades. No granite, schist, or quartzite are known to outcrop in this portion of the range, and it is clear that the erratics were carried through the Gorge from regions east of these mountains. If Diller was aware of the publications of Symons and of Russell on Lake Lewis, he did not suggest correlation between this lake and Willamette Sound.

Here, then, is a curious situation with regard to nomenclature. Condon (1871) has priority of publication, but, so far as we can learn, his conclusion cannot stand on his own presented evidence. It is purely a coincidence that there *was* such a submergence as he names. Symons (1882) has a case nearly identical. If Symons had indicated that he found *erratic* or *glaciated* bowlders, his case would have been clear. Russell (1893) did not know of Condon's publication, or he would have mentioned it when he admitted that the cause of the ponding was not known. He preferred the hypothesis of a glacial dam. It seems unlikely that he did not know of Diller's discovery of granite erratics in the Willamette Valley before his own last-published mention of Lake Lewis (1898). Yet he consistently referred to the ponding as Lake Lewis, clearly not correlating the episode in central Washington with that in western Oregon. Russell and Diller independently discovered the evidence which establishes the existence of the water body under discussion. They are to be credited accordingly, though neither saw as broad a vision as did Condon.

There was never a glacial dam across the Columbia in the Gorge. The drifted bowlders in Willamette Valley, derived from

the east side of the range, are clear evidence of open waters through the Gorge at the time of the submergence. Further, a study of the various tributary valleys along the Gorge shows that local glaciers, instead of filling it to a depth of more than a thousand feet, probably never reached the Columbia. In Hood River Valley there are moraine deposits apparently recording two different glaciations, during the earlier of which the ice reached within 3 miles of the Columbia, and during the later not nearer than 5 miles. No other local glacier got as near as this to the Columbia. It seems impossible that Hood River Glacier could have filled the Columbia Valley to a depth exceeding a thousand feet and left no trace other than these minor deposits back in the tributary valley. Both Condon[1] and LeConte[2] have referred to glacial deposits near The Dalles, but the writer has never seen anything in this region which could be so interpreted. Topographic situations, altitudes, and the location on the lee side of a mountain range all contribute to weaken the hypothesis of glaciation at or near The Dalles. The Yakima Glacier, on the east slope of the range, 100 miles farther north, did not descend below 2,000 feet A.T., though it was fed from mountains much loftier than those near The Dalles.

NEW EVIDENCE ESTABLISHING THE FACT AND EXTENT OF THE SUBMERGENCE

Foreign bowlders and débris.—This can best take the form of a list of occurrences and altitudes of foreign materials along the Columbia and in some of its tributary valleys.

Kelso, Wash. (5 mi. west of). Granite bowlder, 2½ ft. in maximum diameter. 50 ft. A.T. Has a pegmatite dike in it. Dike projects ½ in. above the pitted surface of the granite.

Manor, Wash. (8 mi. north of Vancouver, Wash.). Granite bowlder in schoolyard; angular, with fresh surface. Altitude close to 300 ft. A.T.

Camas, Wash. Granite bowlder formerly on terrace back of the town, at an altitude of about 175 ft. A.T. Has been broken up and used for foundations. A fragment of it was examined.

Willamette, Ore. (4 mi. southwest of). Several granite and quartzite bowlders lying at the western foot of Pete's Mountain at 300 ft. A.T. Very slightly weathered.

[1] Thomas Condon, *The Two Islands* (Portland, 1902).

[2] Joseph LeConte, "Age of the Cascade Mountains," *Amer. Jour. Sci.*, VII (1874), 167.

Canby, Ore. (4 mi. east of). Several erratic bowlders; granite or granodiorite, porphyritic rhyolite, and quartzite. Altitude 323 ft. A.T. Have been blasted out of the soil in clearing a field. Local authority states that a granite bowlder like those in this group lies on the plateau tract 4 to 5 mi. north of Canby. Its altitude must be about 450 ft. A.T.

Dilley, Amity, and Corvallis, Ore. Diller found foreign bowlders near these places in the Willamette Valley, the highest at 250 ft. A.T. He quotes local authority for their occurrence as high as 1,500 ft. A.T. "near" Oswego, Ore., but there are no elevations of more than 950 ft. A.T. near enough to this place to be referred to its locality.

Granite bowlders are very rare in the Gorge, east of the Willamette broadening. Only one has ever been seen by the writer and none are noted in the literature. The lone bowlder lies in Hood River Valley, Oregon, on the summit of Van Horn Butte, a small cinder cone three-quarters of a mile southeast of Van Horn. It is of granite. Its altitude is between 800 and 900 feet A.T. Immediately east of the Gorge, however, the foreign bowlders are more common than in the Willamette Valley. They occur on the eastern flank of the easternmost anticline of the Cascade Range, about the city of The Dalles, Oregon.

The Dalles, Ore. Granite bowlder on summit of terrace immediately south of the city reservoir.

The Dalles. Valley of Mill Creek, 1½ mi. from town. Large granite bowlder lies in the valley; probably has fallen from the summit of the bluffs, about 500 ft. A.T.

The Dalles (about 6 mi. west of), on old Hood River road. A dozen or more angular fragments of granite, 2–3 ft. in maximum diameter, grouped as though the result of frost action or blasting of an original large bowlder. Altitude about 1,200 ft. A.T.

The Dalles (about 6 mi. west of), north of old Hood River road. Granite bowlder, maximum diameter about 3 ft. 500 ft. A.T.

Arlington, Ore. (1 mi. east of). 30 to 40 angular pieces of granite, from 2 to 6 ft. in maximum diameter, grouped within a radius of 25 ft. Altitude 800 ft.+ A.T. Estimated volume of exposed granite material, 500 cu. ft., equivalent to an original bowlder about 7×7×8. The granite is light gray in color, composed of quartz, feldspar, mica, and hornblende, and contains lenses and fragments of a dark ferromagnesian phanerite. Portions are a true intrusion breccia. Many of these huge granite bowlders are said to lie on the slopes south of Arlington. Many of them have been broken up for use in cemetery monuments.

Roosevelt, Wash. Slates, diorites, quartzites, and several varieties of granite are all represented in cobbly and bowldery fragments up to 1,000 ft.

A.T. on the slopes of Wood Gulch back of this town. One quartzite, 8 in. in diameter, has glacial striae on one side. These fragments lie below the outcrop of an intra-basalt gravel formation, but quite surely are not derived from it.

Mottinger, Wash. (4 mi. north of). A distinct mound 4 ft. high and 40 ft. in diameter, covered with, and apparently composed of, glacial débris of various kinds, the fragments of various sizes and shapes. It is literally a patch of glacial till deposited on this unglaciated slope. Altitude about 1,100 ft. A.T.

FIG. 1.—Erratic near Prosser, Wash.

Kennewick, Wash. (3½ mi. south of). Foreign material very abundant on slopes and summit of the ridge. Summit is 900 ft. A.T. Granite, gneiss, quartzite, vein quartz, slate, porphyritic phanerites and a variety of metamorphic rock fragments are present. Striated faces are common. In places here the soil looks like the washed surface of a stony glacial till, though the region is many miles from the nearest glaciated area.

Prosser, Wash. (north of). Slopes below about 1,000 ft. bear many bowlders and fragments of granite and quartzite. The quartzite is prevailingly pinkish to lavender to blue in color, a feature common to most of the angular quartzite in this category. One quartzite is marked with two sets of striae crossing each other. One large granite bowlder, 6×7×10, lies at 840 ft. A.T. along the road. It is almost identical in character with the material at Arlington. Riven fragments lying about the bowlder are equal in volume to half the main bowlder (Fig. 1).

Mabton, Wash. (southwest of). Foreign bowlders and cobbles of granite, quartzite, slate, and basalt in considerable numbers lie scattered over the

surface of the fine, light-colored silt which floors this part of the Yakima Valley. Upper limit 1,200 ft. A.T. These erratics occur in groups, several kinds of rock in each group. Both rounded and angular, and both large and small fragments occur in every group.

Snipes Mountain, Wash., between Granger and Sunnyside. Foreign cobbles and bowlders common, scattered over the hillside up to nearly 1,200 ft. A.T. One granite, 2×3×4, at 950 ft. A.T., has a flat, beautifully striated face, 3×4 ft., undoubtedly a glaciated surface.

Toppenish Ridge and Satus Valley, Wash., 5 to 8 mi. south of Toppenish. A striated quartzite bowlder, associated with granite and other foreign material, at about 1,100 ft. A.T. A granite bowlder at 1,100 ft. on the south side of the Satus Valley.

Ahtanum Ridge, Wash. (10 mi. S.S.W. of Yakima). Large numbers of angular granite and quartzite bowlders and cobbles on the southern slopes, below 1,200 ft., and on the floor of the structural valley immediately south. Grouping of angular fragments indicate that some original bowlders were 4 to 5 ft. in diameter. Material fresh, affected only by frost action and temperature changes. One bowlder along road, about midway between Yakima and White Swan, is clearly smoothed and striated on one face.[1]

Dry Creek Valley, Wash. Glacial material plentiful; granite predominant, but quartzite, diorite, slate, etc., present. Observed upper limit about 1,100 ft. A.T. Grouped in the familiar fashion: large and small, rounded and angular, all together except where slope wash has strung them out. Groups separated by intervals of hundreds or thousands of feet where no débris of this sort is to be found. Glaciated surfaces on some bowlders.

Cold Creek Valley, Wash. Situation similar to that in Dry Creek Valley 1,130 ft. maximum altitude observed.

Priest Rapids, Wash. (4 mi. northwest of). Granite bowlders lie on the slopes of the Columbia Valley up to 1,175 ft. A.T. The highest found is 3×4×6 exposed, probably larger. Several nearly as large at 1,150 ft. A.T. on the same hill. Granite is like that in the Arlington bowlder.

Quincy and Burke, Wash. Between these two places 10 or 12 foreign bowlders were seen at altitudes above 1,200 ft. Two are described as follows: (1) six mi. south of Quincy; granite bowlder, 4×5×10 exposed dimensions, of the familiar basic igneous rock in a light-colored granite. (2) five mi. south of Quincy; granite bowlder, largely buried, exposed portion 12 ft. long; the largest and the highest (1,255 ft. A.T.) found by the writer.

Wenatchee Valley at junction with Columbia Valley. Russell reports "great quantities of angular bowlders, some of them of large size, composed principally of quartzite, granite and schists . . . . about Wenatchee and

[1] George Otis Smith, in *Ellensburg Folio No. 86, U.S. Geol. Surv.* (1903), notes the presence of erratic bowlders in Wide Hollow and Wenas Valley, north of Ahtanum Ridge, and says they "must have been dropped from masses of ice floating in the ponded waters of Yakima Valley."

along the river to the south . . . . which were carried to their present positions by ice."[1] He does not say "floating ice" but his own later work showed that no glacial ice ever reached this place, and these bowlders may be listed with the others floated in the Columbia submergence.

Soap Lake and Adrian, Wash. Erratic bowlders very abundant along the Great Northern Railroad in vicinity of these stations. The railroad here crosses the old Grand Coulee channel of the Columbia, occupied while the Okanogan lobe blocked the main valley to the west. Altitude of Soap Lake station is 1,202 ft., of Adrian, 1,234 ft. Whether these bowlders were deposited from bergs in Lake Lewis or were dropped while en route to the lake along the glacial Columbia was not determined. At any rate, they lie close to the entrance of the river into the lake. The Coulee City terminal moraine is distant 20 to 25 mi. by the Grand Coulee route, 15 to 20 mi. in a straight line, and the agency of glacial ice cannot be called upon to explain their presence here.

Ephrata, Wash. "3.7 mi. south of Ephrata, aluminum tablet in large granite bowlder, 1,283.102." Leveled in 1903, by R. A. Farmer and in 1909 by C. H. Semner, of the United States Geological Survey (note taken from *Bull. 674, U.S. Geol. Surv.*)." This is the highest foreign bowlder ever reported from the bed of Lake Lewis, so far as the writer is aware.

Winchester, Wash. "Twenty feet south of track, opposite section house, in top of granite bowlder, aluminum tablet stamped 1277 T. 1277.333" (data from *Bull. 674, U.S. Geol. Surv.*).

The berg-drifted bowlders in the lower Columbia drainage extend almost, if not quite, from the margin of the Okanogan glacial lobe in central Washington to within fifty miles of the Pacific Ocean. This débris did not come from glaciers on the west side of the Cascade Range, for there are no granite outcrops in the drainage of the Willamette Valley. It did not come from glaciers along the east slope of the same range south of Lake Chelan, for none of the valley glaciers here descended below 1,500 feet A.T. and the submergence did not reach much above 1,250 feet A.T. The Chelan Glacier and the Okanogan Lobe appear to have contributed all the bergs which drifted so widely in this inland sea, and débris which they carried was scattered along 350 miles of the lower Columbia Valley.

Most of this débris lies close to the upper limits of the submergence. This is due to two causes: (1) the opportunity for stranding in shallow water along the shores, where the bergs were

[1] "A Geological Reconnaissance in Central Washington," *U.S. Geol. Surv., Bull. 108* (1893).

held until melting, and (2) the subsequent removal or burial of débris dropped in mid-valleys.

Many of the bergs were derived from basal ice in the parent glacier. The amount of fine material associated with the bowlders, and the striated faces of many bowlders, makes this clear.

Practically all of the débris listed in the foregoing catalogue has been derived from rocks of deep-seated origin, indicating the work of glaciers in the heart of an eroded mountain mass.

The upper limit of the submergence, determined from present data on drifted bowlders, is about 1,200 feet through central Washington. The lower altitudes of the known erratics in Willamette Valley may be because the heavy forest above the cleared bottom land conceals higher bowlders. It may also be due to warping in postglacial time along a hinge line approximately coincident with the axis of the Cascade Range.

Off-shore sediments of the submergence.—It is significant that in the lower valley of Yakima River, Dry Creek, and Cold Creek, considerable silt deposits occur below the upper limits of the erratics. Many square miles are covered by these sediments west of Mabton and south of the Yakima flood plain. Here they are dissected to a subdued badland topography and the erratics are scattered widely over the area: on divides, slopes, and valley bottoms. But the material is incoherent and fails to stand in bluffs, so that one cannot be sure that erratics are *inclosed* in the silt deposit. It is probable, however, that these deposits are correlative with those in White Bluffs, now named the Ringgold formation, and older than the glacial submergence.

It is more significant that investigations on water resources in Quincy Valley[1] have shown the presence of "Pleistocene lake beds" below 1,200 feet A.T. Wells nearly 300 feet deep do not penetrate through these beds on Morrison Flat, a few miles west of Moses Lake. These sediments consist of clay, silt, and sand, with a few beds of gravel, and contain bones and shells. "W. H. Dall states that the fossils collected from the lake beds are fresh-water species, all of which are still living and are not older than the Quaternary period. They are of the boreal type and could have lived in the cold water of a glacial lake."[2] Glacial gravels unconformably

[1] Schwennesen and Meinzer, *op. cit.* [2] *Ibid.*, p. 144.

overlie the lake beds, the gravels having been carried down Grand Coulee from the Okanogan Lobe during the subsiding stages of the submergence.

The situation of these deposits near the ice and at the northern extremity of the submergence, and the presence of intercalated gravelly strata make a strong case for their origin in Lake Lewis. If the interbedded gravels were known to contain material of glacial derivation, the case would be clear. Schwennesen and Meinzer think that this lake was local, and that it discharged to the present valley of the Columbia over two abandoned cataracts south and a little west of Quincy. They apparently do not accept the hypothesis of a Lake Lewis as enlarged and set forth by Russell.

Gravel terraces in the Columbia Valley.—Russell and others have noted the presence of several terraces of gravel at different levels in the Columbia Valley from Chelan upstream. The Great Terrace, as Russell called it, appears to be the highest gravel bench in this part of the Columbia Valley. Its altitude near the downstream termination is 600 to 700 feet above the Columbia according to Russell. The Columbia at the mouth of Chelan River is 670 feet A.T., and the surface of the terrace therefore is 1,270 to 1,370 feet A.T. If the Great Terrace is a delta built in Lake Lewis, as Russell thought, and if the present termination is the original delta front, the submergence stood between 1,270 and 1,370 feet in this part of Washington.

The high line of drifted débris is not known to reach above 1,283 feet A.T., though of course the highest bowlders may not have been found. At any rate the close approximation in altitudes strengthens Russell's case for the delta origin in Lake Lewis of the Great Terrace. If top-set beds on fore-set beds are ever found in this terrace their presence will settle the matter affirmatively, and the altitude of their contact will give the precise upper limit of the submergence in this part of the state. If Russell were correct about this, there should also be deltas correlative with the Great Terrace in such tributary valleys as carried glacially fed, and therefore actively aggrading, streams at this time.

The Columbia Valley contains gravel terraces at intervals from the Great Terrace to Portland. Too little is known of these to attempt their interpretation here, but the upper limit of each

highest terrace of the localities known is lower with distance downstream. Apparently they form a simple system of river terraces. So far as known their surfaces do not bear the drifted erratics. They are believed to have been formed during and after the subsiding stages of the submergence. One set of these terraces, here named the Portland Delta, is described as follows:

The eastern portion of the city of Portland is built on a series of terraces of river gravel, the highest of which is about 300 feet A.T. The 300-foot portions lie east of Mount Tabor and Rocky Butte and extend eastward up the river to the mouth of the Gorge. Considerable tracts at 300 feet A.T. also lie on the Washington side of the Columbia, east of Vancouver. This 300-foot plain has been dissected by subsequent meanderings and braidings of the Columbia, and a large portion of it has been brought down to about 200 feet A.T. One of these abandoned Columbia channels is traceable from Rocky Butte in the northeast part of Portland westward to the Willamette in the heart of the city. Another is conspicuous immediately east of Vancouver, Washington, across to the 300-foot plain.

The gravel composing the 300-foot plain and 200-foot terraces is fresh material, discolored only slightly by weathering at the surface, and all of its constituent pebbles are hard. This criterion distinguishes it from terraces of the much older Satsop formation on the Washington side, the altitudes of the lower of which are by coincidence almost the same as the highest of this series.

The gravel is almost everywhere disposed in fore-set beds which dip westward and northwestward. Any gravel pit in Portland shows this, and the long cut of the Oregon-Washington Railroad and Navigation Company in the eastern part of Portland shows it most strikingly.

This deposit of fresh, delta-bedded gravel, with summit level of 275–300 feet A.T., does not extend down the Columbia farther than 10 miles north of Vancouver, Washington. Its outline is that of a broad fan or Greek Δ in the Willamette structural valley, with its apex at the mouth of the gorge. It appears clearly to record static water at this level in the Columbia Valley west of the Cascade Mountains. The absence, so far as known, of the drifted bowlders on its surface, the absence of infiltrated clay in it, and the evidence

from the distributary channels across it of a lowering water body during its formation all suggest that it was deposited after the maximum of the Pleistocene submergence. This may have been either during a pause in the subsidence of the water, or in a later and lesser submergence. The first interpretation is preferred.

THE SUBMERGENCE ELSEWHERE IN WESTERN WASHINGTON

So great a submergence in late Pleistocene time could not have been limited to the lower Columbia drainage. Its duration was so brief, however, that it is not definitely recorded by shore features or lacustrine plains, and had there been no bergs on it the occurrence perhaps and the extent certainly never would have been known.

There appears scant hope, therefore, of finding records of the submergence elsewhere. The only other region in the Pacific Northwest where a similar episode has been reported is the lowland of Puget Sound. During the latest, or Vashon, glaciation of Puget Sound the sea-level was not more than 75 feet above its present altitude.[1] But after retreat of the ice from the lowland, marine waters submerged the northern part of the region to a depth of at least 290 feet A.T.[2] This is recorded by the presence of marine shells in stratified clay overlying the Vashon till sheet.

Erratic bowlders cannot be used for this study in the Puget Sound Valley or its margining slopes because glacial ice has covered the whole region. South of the terminal moraine of the Puget Sound Glacier, however, they would be as valuable a criterion as in the Columbia Valley. But, so far as known, there are no such bowlders in the unglaciated lowlands of southwestern Washington. This may be charged to lack of intensive search for them, or to absence of bergs in such waters.

GENERAL CONSIDERATIONS

The highest known erratic bowlders in Willamette Valley (323 feet A.T.) and the highest known marine shells in Puget Sound (290 feet A.T.) mark the known upper limit of the submergence

[1] J H. Bretz, "Glaciation of the Puget Sound Region," *Wash. Geol. Surv. Bull. 8* (1913), p. 229.

[2] *Ibid.*, p. 233.

west of the Cascades. The hypothesis that submergence was relatively less on this side of the range may therefore be entertained.

Contemporaneity of submergence and glaciation is clearly recorded east of the mountains. In contrast to this it seems clear that in Puget Sound maximum submergence lagged much behind maximum glaciation, highest water-levels not being reached until the ice front had retreated at least 70 miles and probably more than 100 miles.

Though the tides are felt during lower river stages 150 miles up the Columbia, yet Astoria, a few miles inside the Columbia bar, has a fresh-water harbor. The great volume of the Columbia, then as now, doubtless maintained the fresh-water character of the entire submergence. The downstream journey of hundreds of miles by bergs also indicates a current sufficient to offset the influence of the prevailing westerly winds on these bergs. This fresh-water character of the Columbia submergence explains the absence of marine shells in Willamette Valley such as those deposited in Puget Sound during the same epoch.

From the positive evidence of stranded bowlders the submergence reached a known maximum of 1,283 feet A.T. close to the Okanogan Lobe. Its known maximum at The Dalles was 1,200 feet A.T. The Dalles is 125 miles south of the Okanogan Lobe. On the west side of the Cascades the highest records of the submergence in Puget Sound are 290 feet A.T., and bowlders at 250 feet are reported near Corvallis, Oregon, 160 miles south of the front of the Puget Sound Lobe.

Thus there is no evidence in Oregon and Washington for warping along lines parallel to the general front of the Cordilleran ice sheet, nor in central Washington, for progressive submergence of the glaciated area during retreat of the ice. In fact the entire submergence, save in Puget Sound Valley, was confined to territory beyond the limits of glaciation.

Isostatic depression 125 to 150 miles beyond the extremity of the two most extended lobes, and this without detectible warping, seems an impossibility. Furthermore, the apparent warping along a north-south hinge line in the Cascade Range is wholly out of harmony with isostatic adjustment during deglaciation. The submergence, therefore, is referred to diastrophic movements of greater

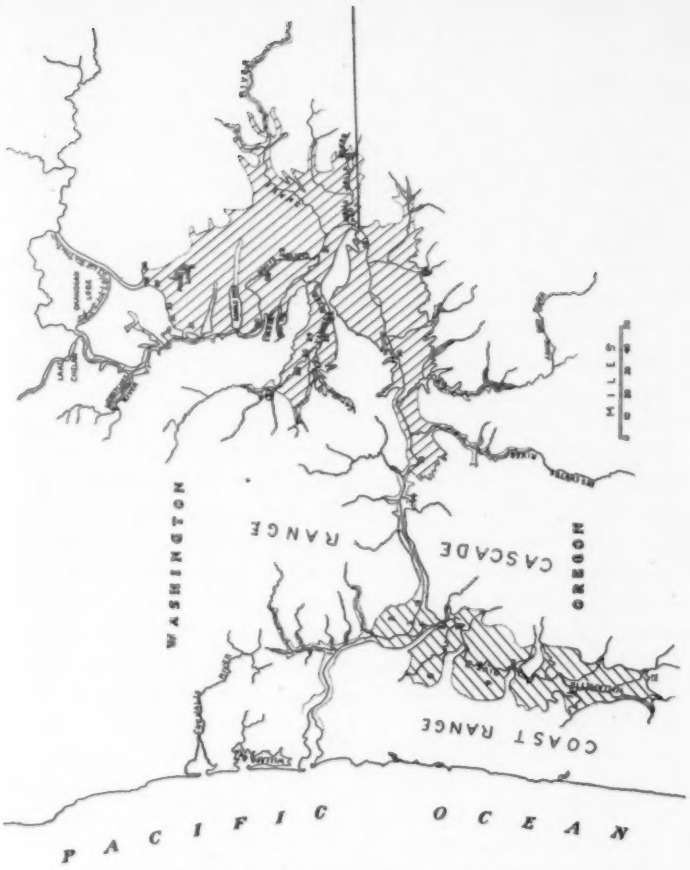


FIG. 2.—The late Pleistocene Columbia submergence

Upper limits of the submergence as shown on this map are 1,250 feet A.T. east of the Cascade Range, 900 feet A.T. in the Columbia Gorge through this range, and 400 feet A.T. west of the range.

N.W.–S.E. diagonal ruling indicates area of Lake Lewis.
N.E.–S.W. diagonal ruling indicates area of Willamette Sound.

1 Kelso, Wash.
2 Manor, Wash.
3 Vancouver, Wash.
4 Portland, Ore.
5 Oswego, Ore.
6 Dilley, Ore.
7 Canby, Ore.
8 Willamette, Ore.
9 Amity, Ore.
10 Salem, Ore.
11 Corvallis, Ore.
12 Eugene, Ore.
13 Camas, Wash.
14 Van Horn, Ore.
15 The Dalles, Ore.
16 Arlington, Ore.
17 Roosevelt, Wash.
18 Mottinger, Wash.
19 Walla Walla, Wash.
20 Wallula, Wash.
21 Kennewick, Wash.
22 Prosser, Wash.
23 Mabton, Wash.
24 Sunnyside, Wash.
25 Granger, Wash.
26 Toppenish, Wash.
27 Yakima, Wash.
28 Hanford, Wash.
29 Ringgold, Wash.
30 Priest Rapids, Wash.
31 Burke, Wash.
32 Quincy, Wash.
33 Winchester, Wash.
34 Ephrata, Wash.
35 Soap Lake, Wash.
36 Adrian, Wash.

extent than, and different genesis from, those resulting from the weight of an ice sheet. In this respect it differs notably from the Champlain submergence of northeastern America.

This hypothesis of a hinge line along the Cascade axis can be most satisfactorily framed in terms of lesser re-elevation west of the range, after the submergence. It is strengthened by the existence of the drowned and alluvium-filled stream valleys of Puget Sound,[1] of the Columbia west of Portland, and of the smaller Chehalis and Willapa valleys. The Columbia east of the Cascades is on a rock floor in many places as far downstream as The Dalles, Oregon.

On the map of the Columbia submergence (Fig. 2) Willamette Sound is assumed to have reached a maximum height of 400 feet. Enough of the valley has been topographically mapped to show that in such a submergence a narrow strait must have existed just south of Portland, and probably another at Salem. Through these straits all the Willamette Valley drainage moved northward.

Diller reports foreign bowlders about Corvallis, south of both straits. It seems impossible that bergs in the Columbia Valley—could have been drifted southward through both straits against the northward flow which must have existed in them. This situation is very strongly suggestive of a submergence in the Willamette Valley even greater than that shown in Figure 2.

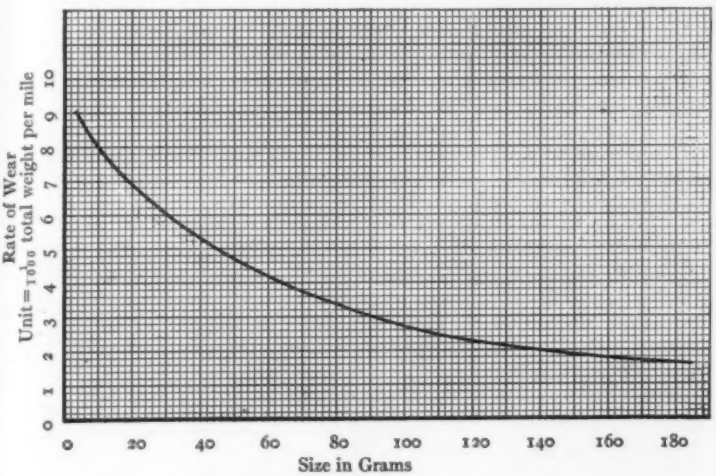
When one plots the approximate extent of this submergence in the Columbia Valley on a map, he finds that there were two broadened portions, one on each side of the Cascade Range, connected by a narrow strait in the gorge, and the whole joined to the Pacific by another strait across the Coast Range. If we were dealing with an existing water body of such outline and magnitude we would almost surely find different names applied to such portions. It is therefore proposed to redefine the names "Willamette Sound" and "Lake Lewis," as outlined in this paper, and to continue their use; to recognize the existence of two elongated straits, one across the Cascade Range, the other across the Coast Range; and to consider the whole as "The Columbia Submergence" of the Champlain epoch.

[1] J H. Bretz, "Glaciation of the Puget Sound Region," *Wash. Geol. Surv. Bull. 8* (1913), chap. viii.

Insert to precede page 507

ERRATA

In Mr. Wentworth's article in the *Journal of Geology*, Volume XXVII, pages 507–21, a number of figures were incomplete and others not properly arranged. For the corresponding figures please substitute the following:



This curve is under the following conditions:

Rock—Niagara limestone
Barrel Vel.—27 R.P.M.
Barrel Diam.—24 inches
Flushed with water
Mixture—3,000 grams in 6½-inch compartment

FIG. 2.—Relation of rate of wear to size of cobble

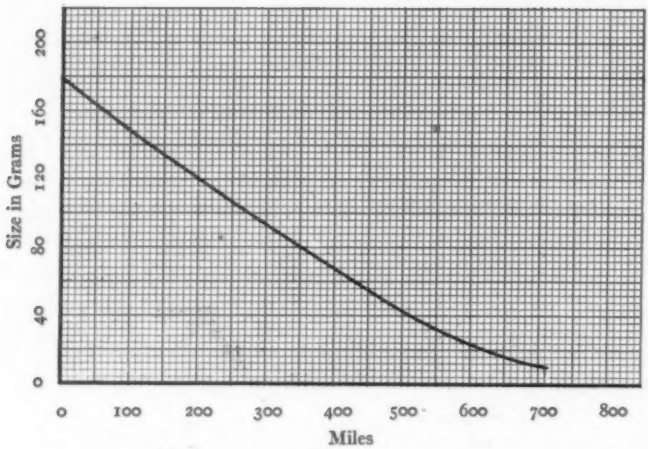


FIG. 3.—Relation of size to distance traveled. Ideal history of cobble starting at 178 grams' weight.

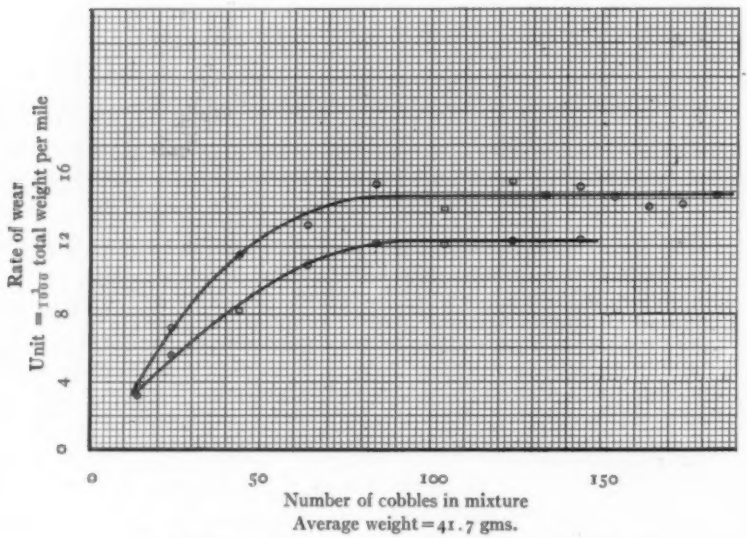
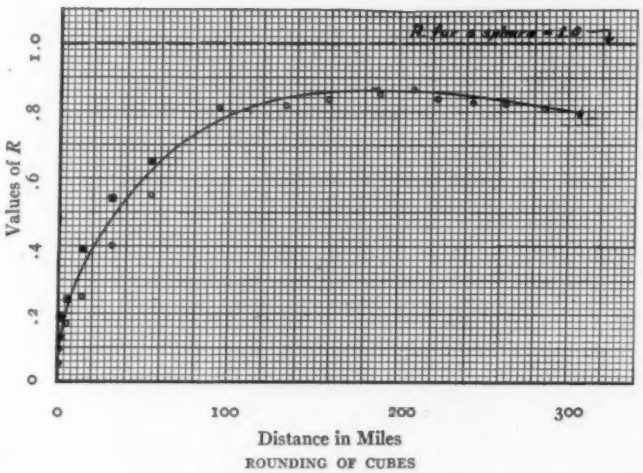


FIG. 5.—Effect of amount of mixture in compartment of drum



ROUNDING OF CUBES

FIG. 26.—Graph of values of *R* plotted against distance for the series illustrated in Figures 8 to 25.

r = radius of curvature of most convex point
d = diameter through same point

$$R = \frac{2r}{d}$$

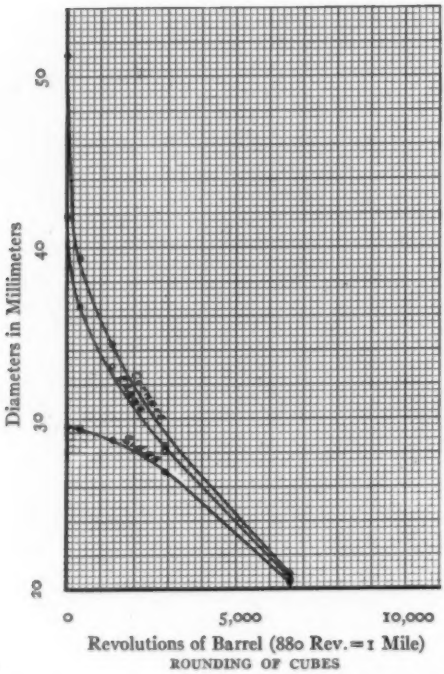


FIG. 27.—Showing convergence of diameters of cubes rounded by cobbles averaging 70 grams each.

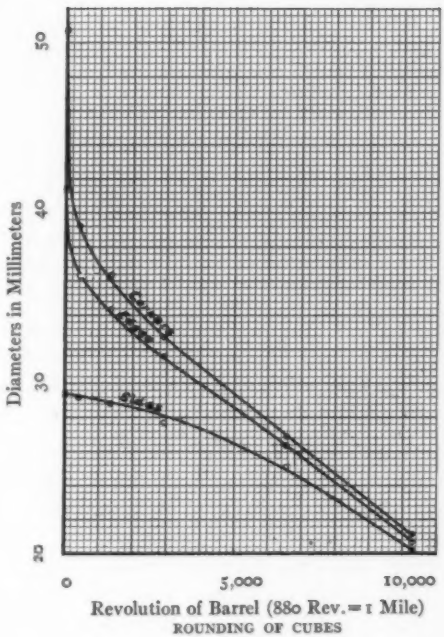


FIG. 28.—Showing convergence of diameters of cubes rounded by cobbles averaging 20 grams each.

A LABORATORY AND FIELD STUDY OF COBBLE ABRASION

Preliminary Report

CHESTER K. WENTWORTH
University of Chicago

INTRODUCTION AND OBJECT

In the course of field work at Baraboo, Wisconsin, in the summer of 1916, it occurred to me that by means of a suitably designed tumbling barrel an experimental study might be made of the abrasion of cobbles in stream transit. It was believed that by carefully controlling conditions, quantitative determination of change of shape and size as related to distance of travel could be made which, when later combined with field studies, would serve as a valuable criterion in the study of deposits of transported materials. Later in in the summer of 1917, during occasional examinations of gravel deposits, the idea was so cordially received by Mr. A. J. Collier, of the U.S. Geological Survey, that I undertook the study of cobble abrasion at the University of Chicago the following autumn and have continued it, with some interruptions, to the spring of 1919. The following is a brief statement of the principal phases of the problem, the methods used, the more significant results obtained to date, and the points which I hope to take up in the future. More complete description of the methods and more thorough discussion of various conditions and their significance will be reserved for a later paper.

OUTLINE OF STUDY

Aside from lithologic character, the distinguishing features of any cobble are its size, shape, and surface texture. These are subject to change and the change is frequently characteristic of the transporting agent.

In stream transportation the rate of change in size of a given piece depends on: (1) its own size; (2) the size of associated cobbles;

(3) the number of such associated cobbles; (4) its angularity; (5) the violence of the motion; (6) the kind of rock (this factor might be subdivided into *elasticity*, *hardness*, etc.). The change of shape depends on: (7) distance traveled; (8) kind of rock; (9) violence of motion; (10) kind of motion; (11) size of cobble; (12) size of associated cobbles. The change of surface texture depends on: (13) kind of rock; (14) violence of motion; (15) kind of motion; (16) size of cobble; (17) size of associated cobbles.

APPARATUS

The apparatus employed consists of two metal drums lined with soft wood, 25½ inches in internal diameter and 13 inches long. These

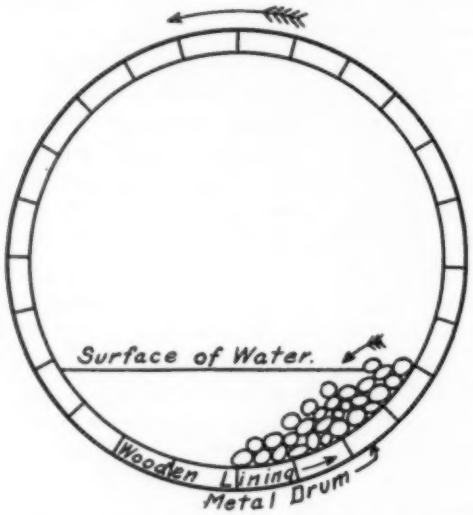


FIG. 1.—Diagram showing in cross-section relations of cobbles to water and to lining of drum while in motion.

are situated at the two ends of the supporting axis and, for all the work to date, have been driven at 27 revolutions per minute. The drums are kept flushed with water so that the finer detritus up to the size of coarse sand is carried out and collects in the settling tubs.

The cobbles in the drums roll continuously down the upgoing side of the barrel, practically under water, at its rate of peripheral travel (1.84 miles per hour or 44.16 miles per day).

The question of how closely the rolling of cobbles down this incline simulates the actual travel of similar materials in streams

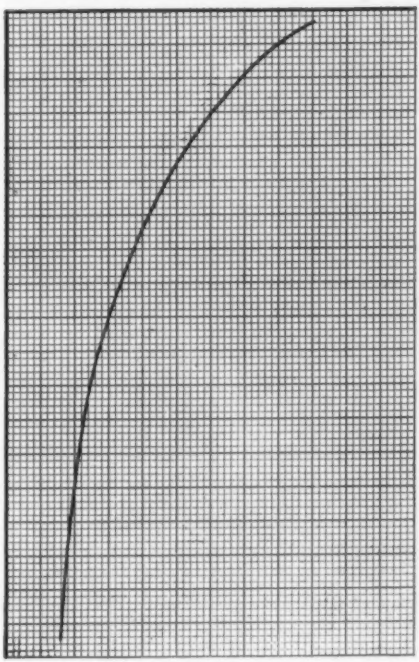


FIG. 2.—Relation of rate of wear to size of cobble

has not yet been carefully investigated. The chief difference, however, will be a constant difference in rate of wear, i.e., 1 mile in a drum equals more or less than 1 mile in the stream, and will not detract from its value in making comparative studies.

It is planned later to make experimental studies of wear during actual transit in streams under natural conditions and

thus determine a factor of comparison to render applicable to field interpretation the comparative results obtained in the laboratory.

I hope also to experiment at length with different speeds and different kinds of motion both in the laboratory and in the field in the hope of throwing some light on the various shapes of cobbles

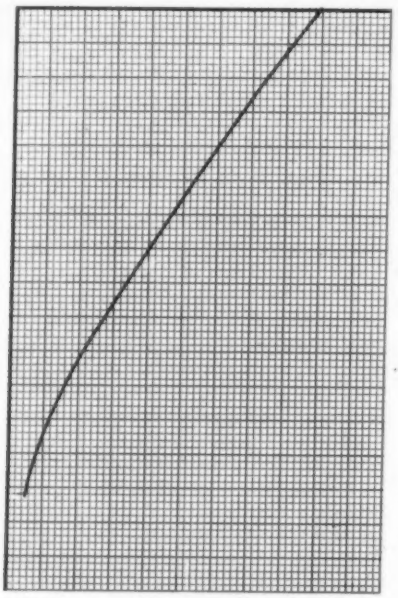


FIG. 3.—Relation of size to distance traveled. Ideal history of cobble starting at 178 grams weight.

thought by some to be characteristic of river deposits, shore deposits, etc. The following are data on some of the relations mentioned above.

RESULTS OBTAINED AND FUTURE WORK PLANNED

1. *Relation of rate of wear to size of cobble.*—I have taken the rate as $\frac{W-W'}{WD}$ where W = initial weight, W' = weight after test, and D = distance of travel, i.e., percentage loss per mile of travel.

Figure 2 shows the plotted average of 85 measurements of rate with cobbles ranging from less than 4 grams to 185 grams. The graph shows very plainly how much more severe the wear becomes as any given piece becomes smaller. This is probably in part due to the increasing ratio of surface to weight.

Figure 3 shows the size history of a cobble started at 180 grams. The rock used in this and all other determinations, unless

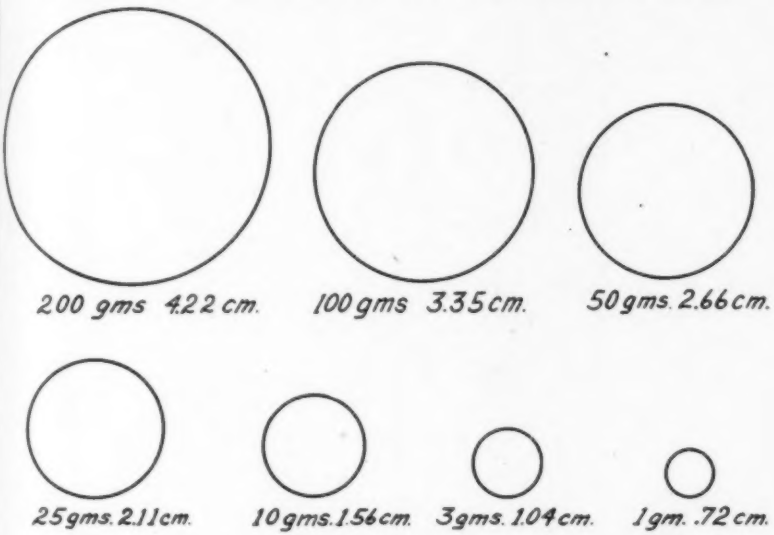


FIG. 4.—Showing sizes of spheres of limestone (density 2.65) of different weights. Weights in grams and diameters in centimeters.

otherwise noted, was Niagara limestone from the outwash gravel at Cary, Illinois. It will be noted that 700 miles of travel reduced the cobbles from 180 to about 10 grams. To aid in visualizing these sizes, Figure 4 may be consulted.

No extensive comparisons have yet been made between different kinds of rock, but measurements were made on one granite cobble which was more than ten times as resistant as the Niagara limestone, and some limestone has been used which was less than one-tenth as durable, so that there is known to be more than one hundred fold

variation in rocks of different sorts. The curve of size history would therefore be notably different in the scale along the abscissa, though the form would be similar for different rocks.

2. *Relation of wear to size of associated cobbles.*—Not yet studied.

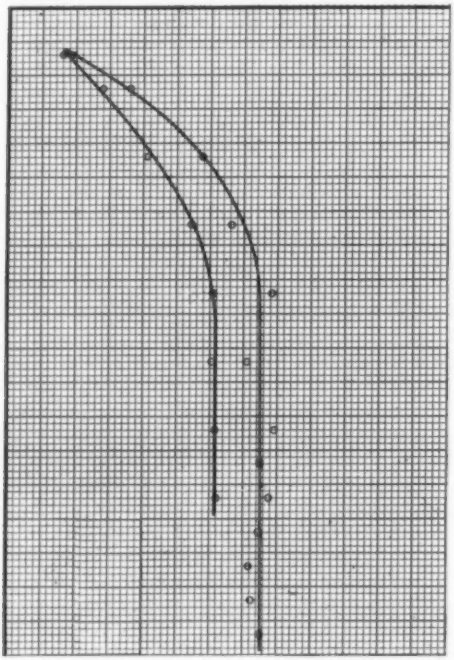


FIG. 5.—Effect of amount of mixture in compartment of drum

3. *Relation of wear to number of associated cobbles.*—That the total amount of material in a compartment of the drum was a very critical factor in the rate of wear was recognized from the first, and to aid in interpreting and correcting for unavoidable variations in the amount, series of runs were made with varying numbers of cobbles in the drum and measurements made on a number of test pieces. The results are shown in Figure 5.

It is to be noted that severity of wear is nearly proportional to the number of cobbles up to a certain point and then remains at an almost constant value. This point, beyond which further additions make little change in rate, is reached when the cobbles are rolling close-packed across the full width of the drum and additional cobbles roll down on top, forming a second layer. Future work will advantageously be done with a filling equal to or greater than 100×41.7 grams to eliminate to a large extent the critical effect of small changes in amount of mixture. The run shown in Figure 5 was started at 144 cobbles, reduced by stages to 14 cobbles, and then increased in an attempt to duplicate the curve. The form is nicely duplicated, while the change in absolute value is due to the diminution in the meantime of the test pieces and their consequent higher rate (see Figure 2); therefore the greater ordinates of the upper and later curve.

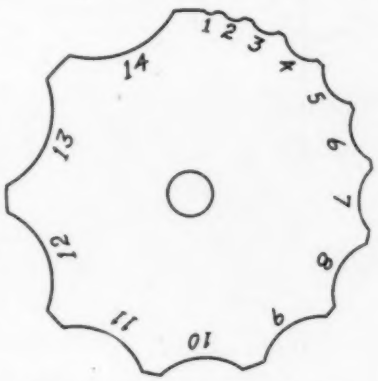


FIG. 6.—Gauge used in measuring curvature. Figures give radii in millimeters.

4. *Relation of wear to angularity.*—Not yet studied.

5. *Relation of wear to violence of motion.*—Not yet studied.

6. *Relation of wear to kind of rock.*—Not yet studied.

7. *Relation of change of shape to distance traveled.*—In stream-transported materials the change in shape, provided the rock is homogeneous, is an approach to sphericity. In studying this change in shape it is necessary to recognize varying degrees of roundness, in other words, to have a numerical answer to the question of how round a given piece is. At least three criteria of roundness readily occur to one in considering the question. These are (1) the ratio of surface area to volume, (2) the average deviation of diameters from a mean diameter, (3) the average

deviation of convexities from a mean convexity. Each of these values or coefficients of roundness reaches a minimum in the case of a sphere and has a maximum in the case of a line or a plane without volume.

Considerable computing and some experimenting was done to test out these methods for the definition of roundness. None was found entirely satisfactory, especially for field use. In order to

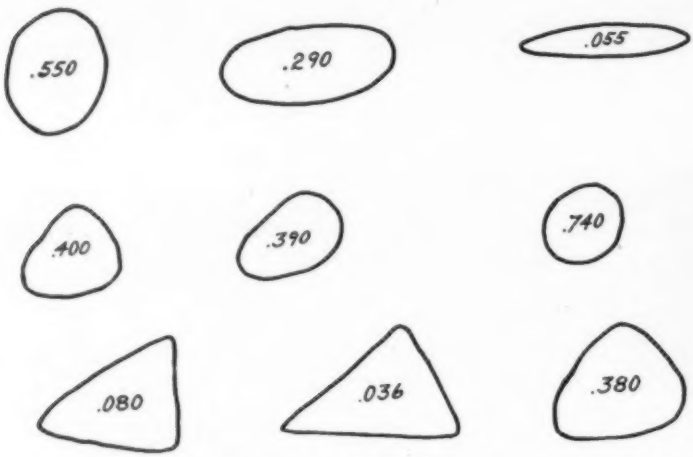


FIG. 7.—Profiles of cobbles showing differing degrees of roundness and giving values of *R*.

be of use in the field study of a deposit, the method of measurement should be simple and rapid enough to permit the calibration of one or two hundred cobbles in a day's time and to allow using the average of the results to characterize the deposit. In (1) the measurement of surface area with sufficient accuracy, while not impossible, was too laborious to use with more than one or two specimens. Likewise in (2) and (3) the need for many measurements of diameters and convexities rendered them liable to the same objection. The method of plotting different kinds of diameters used in the study of cube rounding reported below is of no value for various general shapes. As a practical solution of the

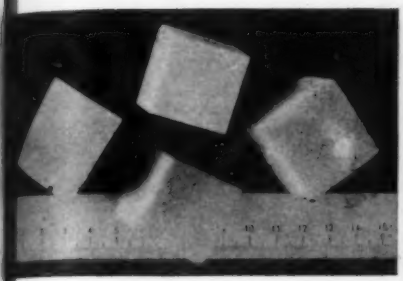


FIG. 8.—Stage 1, 0 miles, $R=0$

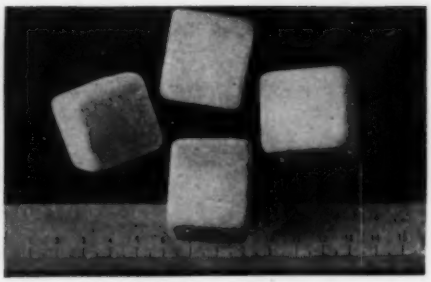


FIG. 11.—Stage 4, 2.18 miles, $R=.130$

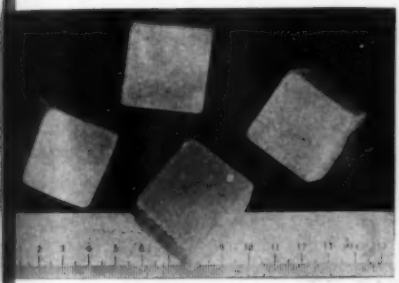


FIG. 9.—Stage 2, .31 miles, $R=.053$

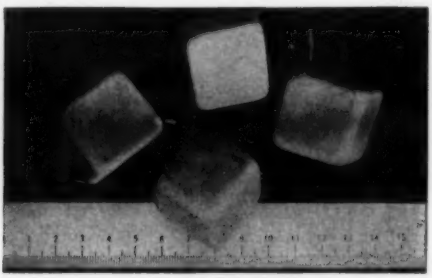


FIG. 12.—Stage 5, 5.12 miles, $R=.170$

FIG. 10.—Stage 3, .91 miles, $R=.096$

FIG. 13.—Stage 6, 14.1 miles, $R=.250$

FIG. 14.—Stage 7, 31.5 miles, $R = .400$

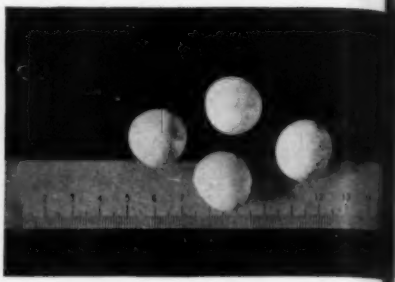


FIG. 17.—Stage 10, 134 miles, $R = .818$

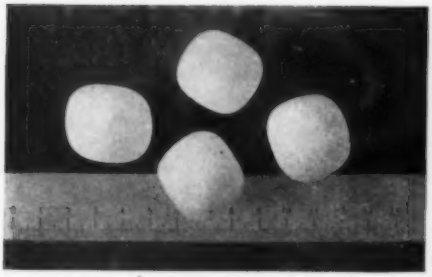


FIG. 15.—Stage 8, 54.5 miles, $R = .550$

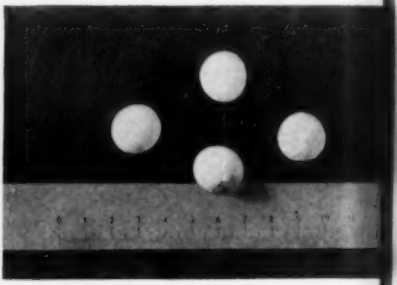


FIG. 18.—Stage 11, 159 miles, $R = .838$

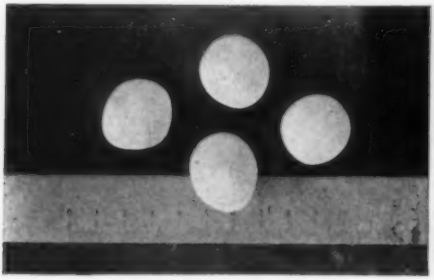


FIG. 16.—Stage 9, 94 miles, $R = .810$

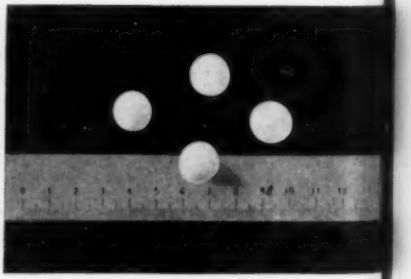


FIG. 19.—Stage 12, 189 miles, $R = .848$

problem, I have used the ratio of the radius of curvature of the most convex part of the surface to half of the longest diameter through that point. This radius is measured with a gauge such as shown in Figure 6 and the diameter with a caliber or usually with a scale with sufficient accuracy.

Figure 7 shows in profile various cobbles with their corresponding values of *R*.

Figures 8–25 show the history of four white marble cubes with the corresponding values of *R*. It will be noted that the four pieces reached a maximum roundness of .862 at an average weight of 3.29 grams and then systematically became somewhat less round as they decreased in size.

Figure 26 is a graph of the roundness as a function of the distance traveled. For this curious increase in angularity with travel I suggest the following interpretation: For any given rock there is

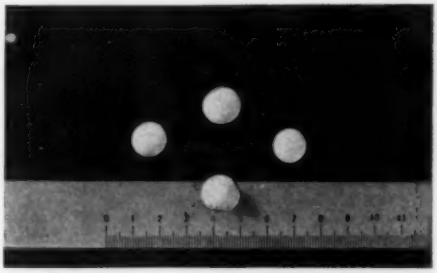


FIG. 20.—Stage 13, 209 miles, $R = .862$

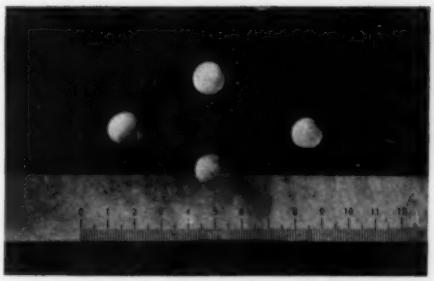


FIG. 21.—Stage 14, 223 miles, $R = .838$

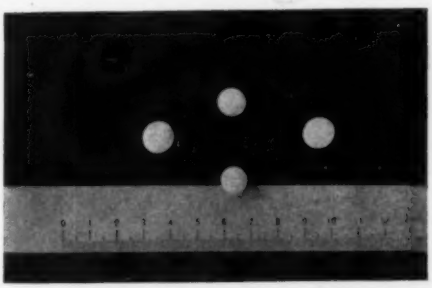


FIG. 22.—Stage 15, 244 miles, $R = .826$

an ellipsoid of equilibrium for wear by abrasion toward which the cobble approaches, which depends on the hardness, or more correctly, durability along various axes.

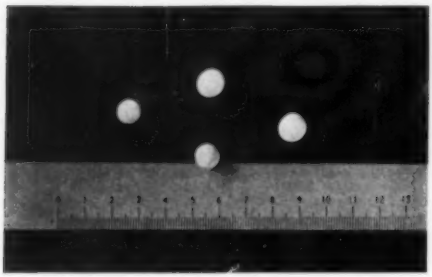


FIG. 23.—Stage 16, 263 miles, $R = .823$

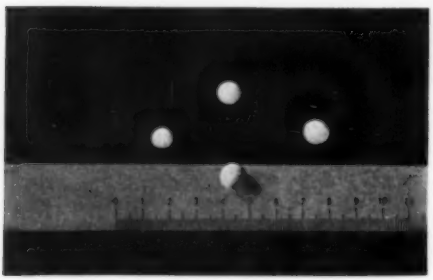


FIG. 24.—Stage 17, 286 miles, $R = .808$

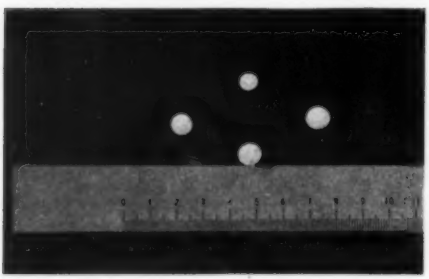


FIG. 25.—Stage 18, 307 miles, $R = .797$

For a perfectly homogeneous rock this ellipsoid is a sphere; for a non-homogeneous rock it is an ellipsoid of greater or less eccentricity as the case may be and has a greater or less value for R in the notation here used. Further, and this is the essential point, this ellipsoid of equilibrium is of different eccentricity for different sizes, more eccentric for smaller sizes than for larger. Thus it is conceived that at 3.29 grams the marble cobbles had nearly or quite reached the ellipsoid of durability for that size and from then on were practically following the equilibrium figure in its decreasing values of R as the sizes grew less. This would mean that slight variations in durability between different directions come out more

notably in small pebbles than in large, which seems reasonable. This interpretation is put forward as a hypothesis which I hope later to test experimentally in considerable detail.

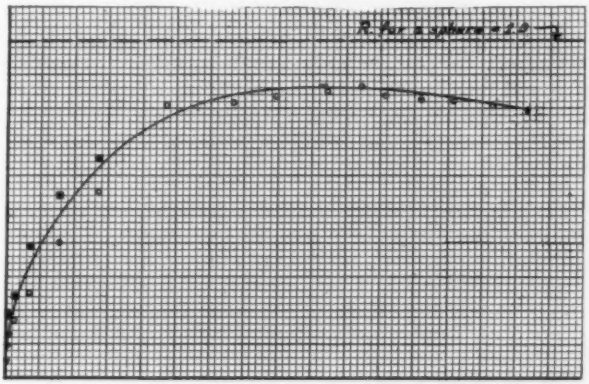


FIG. 26.—Graph of values of R plotted against distance for the series illustrated in Figures 8 to 25.

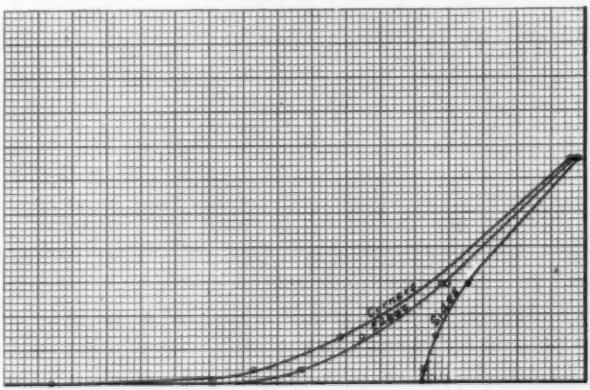


FIG. 27.—Showing convergence of diameters of cubes rounded by cobbles averaging 70 grams each.

8. *Effect of kind of rock on rounding.*—It is hoped later to make studies of the rounding of shaly or schistose rocks whose ellipsoids of

equilibrium would depart notably from spheres, but nothing has yet been done along this line.

9. *Effect of violence of motion on rounding.*—Not yet studied.

10. *Effect of kind of motion on rounding.*—Not yet studied.

11. *Effect of size of cobble on rounding.*—Not yet studied.

12. *Effect of size of associated cobbles on rounding.*—Figures 27 and 28 show the influence of this factor and indicate that large cobbles are much more effective rounding agents than those considerably smaller than the cobble rounded, as shown by the much

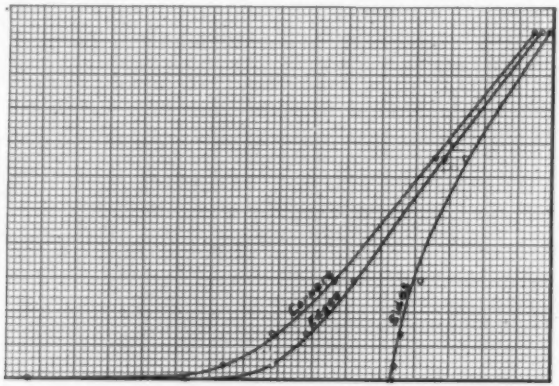


FIG. 28.—Showing convergence of diameters of cubes rounded by cobbles averaging 20 grams each.

slower convergence of the lines in the latter case. It will be noted that the corners and edges lose more rapidly at first (curves concave upward) while the sides lose more rapidly later when their areas are decreased by the wear of corners and edges (curve convex upward). Further studies of this factor are planned.

13. *Effect of kind of rock on surface texture.*—Not yet studied.

14. *Effect of violence of motion on surface texture.*—Not yet studied.

15. *Effect of kind of motion on surface texture.*—Not yet studied.

16. *Effect of size of cobble on surface texture.*—Not yet studied.

17. *Effect of size of associated cobbles on surface texture.*—Not yet studied.

It is hoped that by a study of the crescentic impact scars on some compact quartzites and other rocks an approximate quantitative measure may be made of conditions under which a given deposit was transported and its history thereby interpreted, but no detailed study has been made.

CONCLUSION

Experimental studies, physical or chemical, are of little value to the geologist until they have been applied to actual field conditions. In the studies described and planned above I have tried to keep in mind the possible field applications and hope, as time permits, to point out by field studies the value of each one of the foregoing relations in throwing further light on the past history of the earth. I am especially desirous of receiving opinions on the method of measuring and defining roundness described and used in (17) above, and will appreciate criticism and comment on any other points which have been overlooked or are subject to different interpretation.

CLASSIFICATION OF SPRINGS[1]

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INTRODUCTION

Among the common phenomena of nature, springs are notable because of their high usefulness. Since the earliest times the homes of men have clustered around them. In arid regions their number and size may limit the population. In many humid regions springs are so numerous and similar that distinctions between them are not recognized, yet they may be caused by so many principal and minor factors or by so many combinations of these factors as to make the origin of any one spring exceedingly complex or obscure. An interesting account of the many erroneous

[1] Published by permission of the Director of the U.S. Geological Survey. The illustrations were prepared by the Survey.

notions held by the ancients regarding the origin of springs can be found in Paramelle.[1] In addition, their very familiarity has brought about an indifference which has led the average investigator to pass them by. Only springs with unusual characteristics have been thought worth study. Elaborate classifications have been suggested for so-called mineral springs—those whose water is exceptional because of gas or mineral content—but a complete classification of all springs has been attempted only by Keilhack.[2] His classes are not mutually exclusive, and his primary division into descending (*absteigende*) springs and ascending (*aufsteigende*) springs separates, not waters of unlike origin, but only waters that have unlike paths to the surface. A number of authors have made incomplete classifications for the springs of a limited region or for some special reason. The principles involved and the names used have been helpful in preparing this classification. Such classifications may be found in the works of Gregory, Meinzer, Fuller, and Johnson, cited in this paper, and in those of Hill and Vaughan,[3] Fournier,[4] and Kilian.[5] References to many articles on springs will be found in Meinzer's bibliography of ground water.[6]

The essential factors in the production of springs are the source of the water and the rock structure which brings it to the surface,

[1] M. l'Abbé Paramelle, *L'art de decouvrir les sources*, 4th ed. (Paris, 1896), chap. xi, pp. 64 ff.

[2] K. Keilhack, *Lehrbuch der Grundwasser und Quellenkunde*, Berlin, 1912.

[3] R. T. Hill and T. W. Vaughan, "Geology of Portions of the Edwards Plateau and Rio Grande Plain Adjacent to Austin and San Antonio, Texas," etc., *U.S. Geological Survey, Eighteenth Annual Report*, Part II *b*.

[4] M. E. Fournier, "Etude sur les sources, les resurgences et les nappes aquiferes du Jura Franc Comtois," *Bull. des Services de la Carte Géol. de la France*, etc. Tome XIII (1901–2), No. 89, pp. 1–55, with 31 figs.

M. E. Fournier, "Etudes sur les projets d'alimentation, le captage, la recherche et la protection des eaux potables," *Bull. des Services de la Carte Géol. de la France*, etc. Tome XIV (1902–3), No. 94, pp. 1–30, 1 fig.

[5] W. Kilian, "Essai d'une monographie hydrologique des environs de Gareoult (Var)," *Bull. de Services de la Carte Géol. de la France*, etc. Tome XVI (1904–5), No. 111, pp. 1–9 with 4 plates.

[6] O. E. Meinzer, "Bibliography and Index of the Publications of the United States Geological Survey Relating to Ground Water," *U.S. Geological Survey, Water-Supply Paper 427*, 1918.

and on these factors the classification outlined in this paper has been based. Temperature, dissolved salts, contained gases, rate and amount of flow, form and position of the spring opening are all characteristics of springs, which, while in many cases related to genesis, vary among springs of the same origin. It has seemed best to first divide springs into two groups based on character of the water and make further subdivisions on structural grounds.

In the use of this classification difficulties will arise which are of two types. In the first place, the local structure in the vicinity of springs is difficult to determine, for the presence and passage of water facilitates weathering and destroys the evidence. The presence of luxuriant vegetation also tends to conceal the structure. Whatever the difficulties of determining the structural relations and origin of the water for single springs, the study of groups of springs will usually be successful. The second difficulty arises through various combinations of structures which may combine to produce a spring. The structure which plays the predominating rôle should then determine the classification of the spring. The common sense and judgment of the investigator will lead him to the proper decision, but his labor will be easier if he remembers that ground water moves through three dimensions, though our conventional methods of illustration show but two. Springs of diverse origin may, however, have peculiarities so remarkable or interesting as to justify their grouping under a common name. The proposed system is not intended to supplant the use of such descriptive terms as blowing springs and thermal springs, but to provide a series of terms expressive of genesis which will include all springs, particularly those now called, for want of a better term, "common springs."

ACKNOWLEDGMENTS

In preparing this paper the writer has profited by the kindly criticism of members of the Ground-Water Club of Washington, D.C. The late Professor Joseph Barrell and Professor Herbert E. Gregory offered helpful suggestions. The writer wishes to acknowledge his special obligations to Dr. Henry Hollister Robinson for his generous help in methods of expression.

DEFINITION OF THE TERM "SPRING"

A spring is a place where water issues from the ground and flows or where it lies in pools that are continually replenished from below, except that wholly artificial openings, such as artesian wells, are not regarded as springs. Many springs have been modified by structures intended to increase their usefulness to man. A seep is a variety of spring in which the water comes, not from any definite opening, but through the pores of the ground over a considerable area. The amount of water yielded by most seeps is small. Many marshes and swamps are actually seeps on a large scale. Large ponds or lakes that are supplied with water through openings in their beds are called spring-fed lakes. A series of seeps and springs may occur along a line, which is then called a spring line. Many local names are used for springs, such as "water hole," "ciénega," and, in New Mexico, "ojo." Names like American Water, Bennet's Wells, Coyote Holes, and Ojo de Gato applied to springs show characteristic usage in the arid Southwest.

The conditions and processes that give rise to springs should be distinguished from those that bring about capillary discharge of ground water. The water of springs and seeps rises under pressure transmitted through the water as it lies as a continuous body in the voids of the rock. On the other hand, capillary discharge is due to molecular attraction between the soil particles and the water, acting against gravity. It takes place because the water is raised from the water table through minute openings in the soil by the force of capillarity and evaporates at or near the land surface. No water is released except by evaporation into the air, whereas the water of springs and seeps forms streams and pools unless the quantity is small and the evaporation excessive. The limit of depth to the water table necessary for capillary discharge to be effective is dependent on the size and uniformity of the soil grains. The limit of capillary rise for most soil is not over ten feet. Many areas that fulfil these conditions are large and well defined—for example, the alkaline flats of arid basins.

CHARACTERISTICS OF SPRINGS

In addition to the rate of flow, other characteristics or peculiarities of springs have given rise to names and classes. None of

these systems of classification are complete enough to include all springs, nor are the classes established even mutually exclusive. The names record only the peculiarities of springs, though these peculiarities may and many of them do arise from diverse causes.

Mineral springs are those which yield water containing in solution (1) unusual amounts of mineral matter, or (2) some uncommon or especially noticeable mineral matter. In distinction other springs are called "common springs." Ground water takes up soluble substances from the rocks through which it flows. In consequence small quantities of soluble matter near the point of emergence of spring water are very effective in changing its composition. Thus mineral content is at best an uncertain guide to the origin of the water or the cause of the spring. Usage also is not consistent. Many "mineral springs" yield water of a type that is common in adjacent regions, but because it is unusual in the immediate neighborhood they are distinguished from "common springs." If the water has or is supposed to have therapeutic value, "mineral springs" are often called "medicinal." Mineral springs are classed according to the chemical composition of the water, and one of the most elaborate classifications is that of Peale.[1]

Some of the simpler and more generally used terms are self-explanatory. Saline springs contain common salt; sulphur springs contain compounds of sulphur, usually hydrogen sulphide; chalybeate springs contain iron; calcareous or lime springs contain calcium carbonate; gypsum or "gyp" springs, gypsum; borax springs, borax, etc. Oil springs contain petroleum suspended in drops in the water. The drops of oil usually rise and form a thin iridescent film on the surface of the water. Inflammable gas may accompany the oil or may occur alone in spring water. Such spring waters have risen through or near beds containing petroleum or natural gas. False oil springs also occur. The iridescent film in these springs is due to iron hydroxide, which at one stage in its formation produces the film. In some springs an oily scum is produced by the decomposition of plants or animals buried but a few inches or feet below the spring opening.

[1] A. C. Peale, "The Natural Mineral Waters of the United States," *U.S. Geol. Survey, Fourteenth Annual Report* (1894), Part II, p. 66.

Springs may be divided according to temperature into thermal and non-thermal springs. Most non-thermal springs have temperatures that are approximately the same as the mean annual temperature of the air of the region in which they are found. The division between thermal and non-thermal waters is usually fixed at 70° F., but 20° to 25° above the mean for the region might be preferable. Thermal springs are usually called "hot," but those of slightly lower temperature are sometimes called "warm." Cold springs have temperatures below normal. The water of some cold springs is derived from the melting of ice or snow; that of others, being quickly transferred from a higher to a lower elevation through open channels, retains the temperature of its point of origin on emergence.

Boiling springs have a sandy bottom, through which the water emerges with some force. The sand is constantly agitated and appears to boil. Bubbling springs, also called boiling springs, are due to the emission of gas or vapor with the water. Certain hot springs actually boil in the ordinary sense. Usually emission of air or gases gives the impression of ebullition. Carbonated springs, which emit carbon dioxide, are the most common bubbling springs. Nitrogen, hydrogen sulphide, sulphur dioxide, marsh gas, and other gases have been found in spring waters. Bubbling is produced also by the emergence of water from a well-defined opening under considerable pressure into a pool of water. The surface of the water is domed, and slight fluctuations in volume or pressure give a bubbling effect.

Perennial or permanent springs flow throughout the year. Intermittent or temporary springs flow only during or after rain. Where evaporation is high the flow of springs is much decreased or may cease during the warm season. Some springs flow only at night because of a very delicate adjustment between supply of water and evaporation. Periodic springs flow at full strength for long or short periods, which are not closely related to the fluctuations in rainfall. The periodic action is dependent on the existence of open cavities. Springs here classed as solution tubular or cavern springs are most likely to have this characteristic. A cavity is drained by a small but insufficient outlet. As water

accumulates it may finally overflow through some higher opening and give rise to a periodic spring. Extreme periodicity of flow however, is attained only when the exit tube acts as a siphon so as to drain rapidly the water that has accumulated in the cavity during a considerable time.

Geysers are hot springs which at regular or irregular intervals emit a stream of mingled steam and hot water. The vent from which eruption takes place usually lies at the bottom of a pool of clear water, situated at the top of a conical mound of siliceous sinter. The sinter is deposited from the water in successive sheets of gelatinous silica, through the aid of living algae.

The geyser consists of a tube of hot water extending into the ground. The temperature of the water at the surface is about 212° F., but that of the water below the surface exceeds the normal boiling-point of water. The water in the lower part of the tube is prevented from boiling by the pressure of the overlying column. When the temperature at any point in the tube exceeds the boiling point for that depth, steam is formed. The expansion and rise of steam bubbles cause the water to overflow at the top. The consequent relief of pressure throughout the column of water causes instantaneous formation of steam from the superheated water. The result is an eruption. Of the water thrown out, part is lost and part returns to the tube. The next eruption occurs after the accumulation of sufficient water and an adequate rise in temperature.[1]

The three known geyser regions of the world are in New Zealand, Iceland, and Yellowstone Park, in each of which the geysers are associated with active or relatively recent volcanism. The heat may be attributed with certainty to still uncooled igneous rock. The water, however, may have either a deep-seated or a shallow origin.[2] Geyser action is not dependent on the origin of the water but on the existence of proper channels and the requisite temperatures. It seems likely that the presence of silica in solution is

[1] Bunsen and Descloiseaux, *Compt. Rend.*, XXIII (1846), 934, and other papers quoted by Archibald Geikie, *Textbook of Geology* (New York, 1902), p. 405; also W. H. Hobbs, *Earth Features and Their Meaning* (New York, 1912), p. 193.

[2] Arnold Hague, "The Origin of the Thermal Springs in the Yellowstone National Park," *Geol. Soc. America Bull.*, XXII (1911), 103–22.

essential to the formation of the tubes. Certainly the waters of all known geysers carry silica in solution in relatively large amounts.[1]

Ebbing and flowing springs occur along the seacoast. During high tide the sea water acts as a dam for the ground water and causes it to flow inland and at higher levels as springs. At low tide the ground-water level falls and the springs are reduced in volume or dry up. On this account the flow of springs adjacent to the shore fluctuates with the tide and may even cease at low tide except close to the shore, where some springs flow only at low tide. Rise and fall of water level due to the same cause have been noticed in wells.[2]

Blowing or breathing springs and wells are characterized by the emission of air, often accompanied by a trumpeting sound. They appear to be due to two causes: (1) rise of the water table, which causes the expulsion of air from cracks and pores of the rock;[3] (2) decrease in barometric pressure, which causes a similar movement.[4] The rise of water brings about blowing at relatively long and irregular intervals, but low air pressure recurs in short cycles.

Mound and knoll springs occur in arid climates. The water emerges at or near the top of a mound which has been built up by the accumulation of wind-blown sand and dust in the belt of vegetation surrounding the spring. The height of the mound is limited by the height to which the water can rise, for any accumulation of sand that is not moist is easily removed because it cannot support the protective vegetation. When the water can rise no higher, the process which builds the mound tends to ceil it over. If the water then finds a new and lower outlet, the mound is drained and subjected to erosion, especially by wind. Mound

[1] F. W. Clarke, "The Data of Geochemistry," 3d ed., *U.S. Geol. Survey, Bull. 616* (1916), p. 196.

[2] A. C. Veatch, "Fluctuations of the Water Level in Wells, with Special Reference to Long Island, N. Y.," *U.S. Geol. Survey, Water-Supply Paper 155* (1906), pp. 10 ff., 63–69. H. B. Woodward, *The Geology of Water Supply* (London, 1910), p. 90.

[3] H. B. Woodward, *op. cit.*, pp. 87–88.

[4] A. Strahan, "The Movement of Air in Fissures and the Barometer," *Nature*, Feb. 15, 1883, p. 375, quoted by Woodward. A. C. Veatch, *op. cit.*; C. W. Hall, O. E. Meinzer, and M. L. Fuller, "Geology and Underground Waters of Southern Minnesota," *U.S. Geol. Survey, Water-Supply Paper 256* (1911), p. 90.

springs are found in many arid countries in places where water emerges under pressure. Good examples occur in the Tularosa Basin in New Mexico.[1]

Pool springs have large, deep orifices filled with clear water. The pool is surrounded and partly covered with a shelf of fine earth supported by a network of vegetable fibers. The shelf is formed, like mounds, by the growth of vegetation and the filling of the ensuing tangle with wind-blown sand and dust. The two types occur in adjacent springs of the Fish Springs region, Utah, as described by Meinzer.[2] (See also p. 535 and Fig. 3.)

Many springs deposit around their mouths mineral matter that is carried in solution by the water. Calcium carbonate and silica are the most common minerals of spring deposits. Mounds, platforms, and ridges of considerable size are thus formed.

Mud volcanoes—low conical mounds having a crater at the top through which the water rises—are built up when water containing clay or fine sand rises to the surface under pressure. As the water spreads out of the vent it loses velocity and therefore deposits the matter it had carried in suspension. Such deposits are made by the temporary springs formed along fissures in unconsolidated rocks during earthquakes.[3] The emission of volcanic steam and gas through beds of tuff produces the same topographic form. "Salses," "air volcanoes," and "macculutos" are names applied to such springs in different parts of the world. In many of these springs the gases appear to be due to chemical changes in the earth, rather than to volcanism.[4]

SPRING WATER

The waters which circulate in the ground may be roughly divided into two types: (1) deep-seated waters, and (2) shallow waters. The shallow waters are derived largely from precipitation

[1] O. E. Meinzer, "Geology and Water Resources of Tularosa Basin, New Mexico, *U.S. Geol. Survey, Water-Supply Paper 343* (1915), p. 52.

[2] O. E. Meinzer, "Ground Water in Juab, Millard, and Iron Counties, Utah, *U.S. Geol. Survey, Water-Supply Paper 277* (1911), pp. 44–45.

[3] C. E. Dutton, "The Charleston Earthquake of August 31, 1886," *U.S. Geol. Survey, Ninth Annual Report* (1889), pp. 28–284, Plate XX.

[4] Archibald Geikie, *Textbook of Geology* (New York, 1902), p. 407.

at the surface and move through openings, which are generally supercapillary in size. Their movement is due to gravitative pressure transmitted through a continuous body of water, lying in the pore spaces and fractures of the rocks, i.e., by hydrostatic head. Since both kinds of openings decrease rapidly in number and size below 1,500 feet, these waters are limited in amount below that depth. Deep-seated waters have a complex origin. They doubtless include water derived by absorption from the surface, water entrapped in sedimentary rocks at the time of their deposition, and water expelled during the crystallization of igneous rocks. It is believed that these waters do not move because of hydrostatic head, that is, that they are not connected with any overlying and connecting body of water, but that flow is the result of other agencies operative deep within the earth.

Evidence that a spring water has a deep origin may be positive or negative. Thus the water of a spring that has a strong uniform flow not subject to seasonal changes and a high temperature probably has a deep-seated origin. The minimum depth from which the water may come can be roughly estimated from the temperature of the water, on the assumption that there is 1° F. increase in temperature for every 60 to 100 feet of increase in depth. In volcanic regions, however, the increase may be more rapid.

The presence of important breaks in the earth's crust or of other structures along which water could rise furnishes additional positive evidence. Negative evidence is usually easier to obtain and consists of the absence of any structure which could lead the water from the surface to the necessary depth and then return the water again to the surface.

CLASSIFICATION

I. SPRINGS DUE TO DEEP-SEATED WATER

Springs due to deep-seated water may be divided into two classes, according to their geographic distribution, with respect to localities of volcanic or tectonic disturbance. Their relations to the structure of the upper part of the earth's crust and the probable character of the fissures in the zones of fracture which permit the water to rise are shown diagrammatically in Figure 1. In this

figure, no attempt has been made to show the complicated and closely spaced fractures of the upper part of the crust. Only persistent fissures reach the surface, the others merge into the maze of minor fractures and joints.

Volcanic springs are associated with present or past volcanism. This direct association implies that they have their origin either in water expelled from the underlying magma or in surface water that has come into contact with highly heated rocks and has acquired definite characteristics from this association. In general such springs have strong and relatively constant flows and are highly

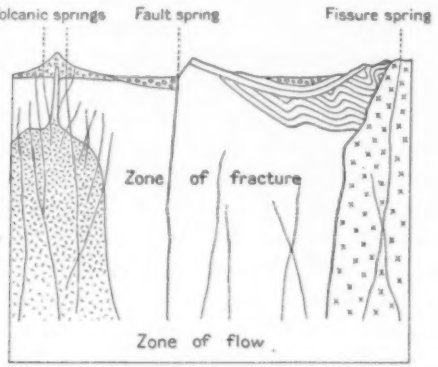


FIG. 1.—Diagram illustrating springs due to deep-seated water

mineralized, and many are gaseous. Some of them have very high temperatures and cannot be sharply divided from volcanic vents emitting steam and gases—fumaroles, solfataras, and mofettes. Others have relatively low temperatures, and many contain carbon dioxide gas. It is generally difficult or impossible to distinguish the latter from springs of non-volcanic origin. The peculiar phenomenon called geyser action characterizes some hot volcanic springs. A typical group of volcanic springs without geyser action lie on the south flank of Lassen Peak, in California (Fig. 2).

The other group of springs due to deep-seated water may be termed fissure springs. In general they have a strong and constant

flow, not subject to annual fluctuations. They are usually warm or hot, and many are highly mineralized. They appear to rise along deep fractures extending far into the crust of the earth. The fractures are similar to those in which were formed the veins now found in mining operations. Doubtless the waters that deposited veins in many places reached the surface as springs.

Certain fissure springs lie along definite lines, and these lines are known to be recent faults involving earth blocks of great depth.

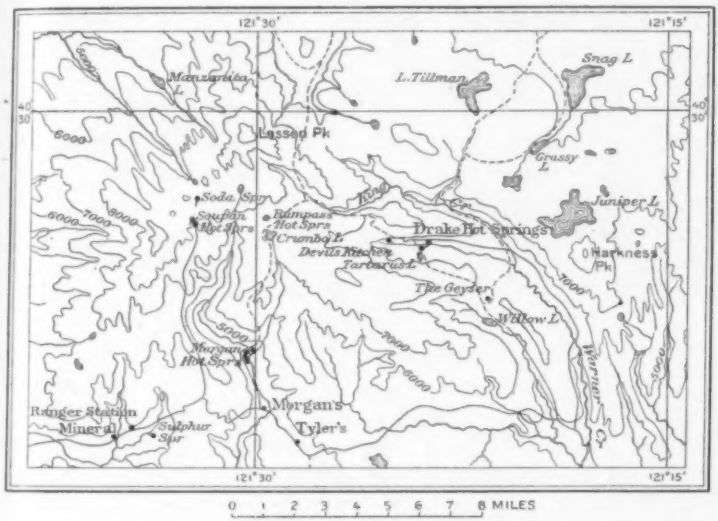


FIG. 2.—Sketch map of Lassen Peak region, California

Faulting appears to produce the fractures which allow these deep waters to rise and carry the temperatures of the deeper crust to the surface. Such springs may be called fault springs. The thermal springs east of the Fish Springs Range, in Juab County, Utah, are a classic example of springs of this type (Fig. 3). The Fish Springs Range is the result of block faulting. The range has a distinct tilt to the west and the fault runs along the eastern flank. Very recent faulting is shown by a fresh fault scarp in the alluvium

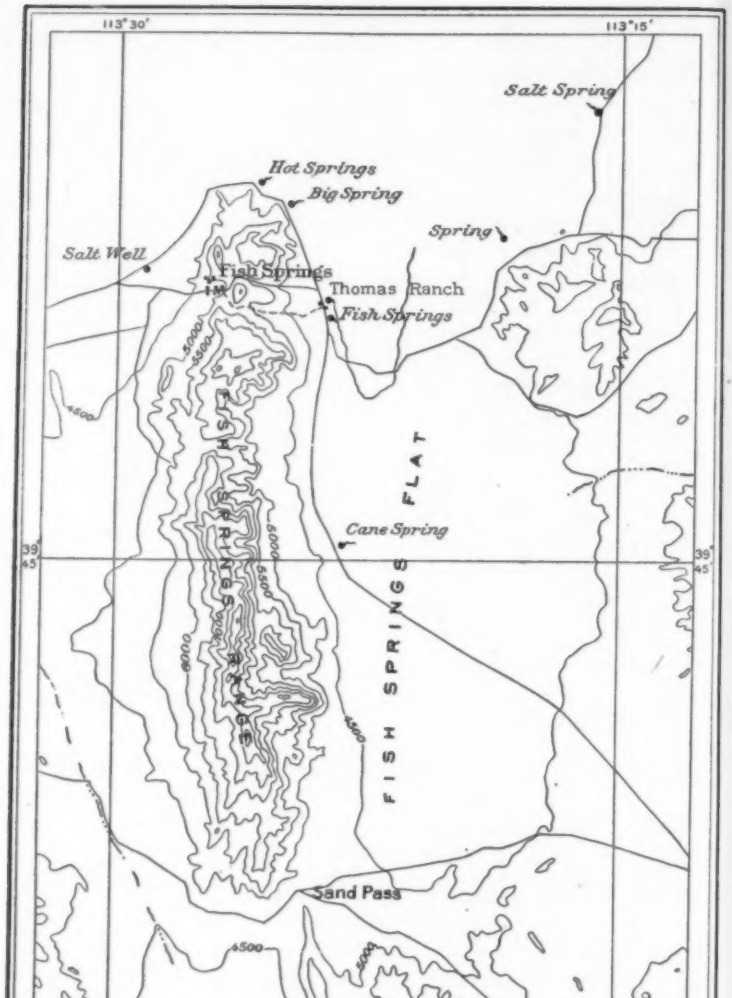


FIG. 3.—Sketch map of Fish Springs Range, Utah

near Fish Springs.[1] Four groups of springs lie east of the range. Hot Springs, Big Springs, and the Fish Springs are on a curving line close to the mountains. They have a steady flow of somewhat mineralized water, and their temperatures are between 78° and 104° F. Cane Spring lies 7 miles south of the Fish Springs group, at the base of the alluvial slope. It consists of a number of seeps of highly mineralized water. Up the slope to the west is the Devil's Hole, a pool of water about 10 or 15 feet below the surface. The form of the pool is similar to those of the Fish Springs group.[2]

Hot Springs, Big Spring, and the Fish Springs are too close to the mountain front to receive much water from the alluvium. Their steady flow and mineralized water imply a deep-seated source, and the recent fault with which they are associated is quite certainly the channel along which they rise. The Cane Spring has the situation of ordinary border springs, but the character of its mineralization and its large volume suggests that part of the water is derived from the now buried southern extension of the same fault. The Devil's Hole may be the last survivor of a group of fault springs whose water has been diverted into the alluvium and rises along the edge of the flat in Cane Spring.

On the other hand, there are fissure springs for whose origin there is no structural evidence. They are believed to have a deep origin because they are not associated with any surface structure that would warrant so strong a flow of water and because of their heated or mineralized condition. Of 98 groups of hot springs in California, 36 occur in granitic rocks.[3] Some of these are fault springs, but others are not directly associated with known faults and can be accounted for only on the supposition that they rise along open fractures or fissures that extend into and draw water from the deeper parts of the earth's crust.

[1] G. K. Gilbert, "Lake Bonneville," *U.S. Geol. Survey, Monograph I* (1890), p. 353.

[2] O. E. Meinzer, "Ground Water in Juab, Millard, and Iron Counties, Utah," *U.S. Geol. Survey, Water-Supply Paper 277* (1911), pp. 43–45, 124–26.

[3] G. A. Waring, "Springs of California," *U.S. Geol. Survey, Water-Supply Paper 338* (1915), p. 154.

II. SPRINGS DUE TO SHALLOW WATER

The pore spaces of the upper crust of the earth are filled with water below a certain level called the water table. This zone of saturation generally has an indefinite extension downward, but in relatively few deep wells is much water encountered below 1,500 feet. Springs due to these relatively shallow waters may be divided into four large groups, according to the character of the rock in which they occur: (1) springs in porous rock; (2) springs in porous rock overlying impervious rock; (3) springs in porous rock between impervious rock; (4) springs in impervious rock.

A. SPRINGS IN POROUS ROCK (DEPRESSION SPRINGS)

Springs in porous rock are formed where the water table or upper surface of the zone of saturation reaches the surface of the ground. Because they are due to the depression of the land surface down to or below the water table, the group may be called depression springs. They are springs whose flow is rather gentle or pools of water that are continually replenished from below. Many are seeps or, if extensive, swamps. They may be divided according to topographic position into four classes, as described in the following paragraphs:

Dimple springs are due to depressions in hillsides which permit the land surface to cut the water table (Fig. 4 *a*). Such depressions or dimples in a sloping surface may arise through erosion by water or wind, through slumping and landslides, through the overturning of trees, or through the operations of burrowing animals, and many are enlarged and deepened by the trampling of larger animals.

Valley springs are due to the abrupt change in slope at the line between the bounding valley walls and the edge of a flood plain (Fig. 4 *b*). Along this line the water table may reach the surface and form seeps or springs. Gullies or low spots between adjacent small alluvial fans may determine the point of emergence. Various causes may enlarge these depressions or may concentrate the flow of the water at specific places, as in the dimple springs.

Channel springs are due to depressions in flood plains or alluvial plains caused by the channel cutting of streams (Fig. 4 *c*). They

thus include all sorts of side channels, abandoned channels, oxbow lakes, sloughs, and water holes. Springs of this type are exceedingly valuable in arid regions. They are frequently made, destroyed, and remade by streams that carry large quantities of sediment.

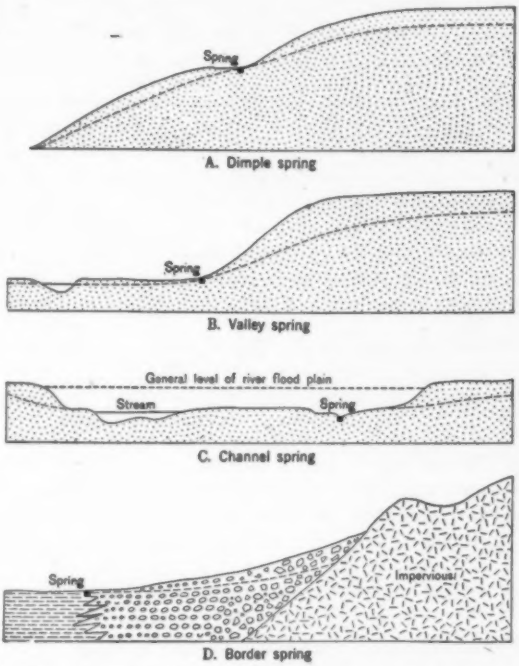


FIG. 4.—Diagram illustrating four classes of depression springs: *a*, dimple springs; *b*, valley springs; *c*, channel springs; *d*, border springs.

Border springs are due to the change in gradient at the line between the alluvial slopes and central flat of a desert basin (Fig. 4 *d*). Such alluvial slopes extend from the mountains with a gradually decreasing gradient. This decreasing slope finally merges with a central undrained flat or with the river bottom of a through-flowing stream. Around the edges of such a flat or the river bottom of a through-flowing stream is a line of springs and seeps.

Ground water is brought to the surface primarily by the change in slope, but the dense silts and clays of the center of the basin tend to act as a dam to prevent further movement down the slope. Thus by their presence they assist the water to rise and may determine the exact location of certain springs. Such border springs are common in the intermontane valleys or bolsons of the western United States. Those of the Big Smoky Valley, Nevada, are shown in Figure 5 and have been described[1] as follows:

> The main west-side spring line of the upper valley extends, with a sinuous course due to the different sizes of alluvial fans, from the Vigus ranch to Wood's ranch, a distance of more than 30 miles, and includes innumerable springs that discharge a part of the copious underground supply received from the Toyabe Range. On the east side of the upper valley there is no spring line comparable to that on the west side, probably because the supply from the Toquima Range is smaller than that from the Toyabe Range, but numerous springs similar to those on the west side are found for a distance of 3 miles in the vicinity of the Charnock ranch.

Concerning the Darrough Hot Springs, which lie along this same line, Meinzer[2] says:

> The water issues from bowldery fill, but probably comes originally from the underlying rock, the heat being due either to igneous intrusion or to faulting that opened deep fissures, or to both causes. Less than 100 feet from the main hot spring and at a level a few feet higher there is a small spring that issues at a temperature of only 60° F., which is almost the normal temperature for this region.

B. SPRINGS IN POROUS ROCK OVERLYING IMPERVIOUS ROCK (CONTACT SPRINGS)

Where porous rock overlies impervious material the water that accumulates in the porous rock is forced to the surface at the contact. Springs so formed may therefore be called contact springs. The form and attitude of the surface of the underlying impervious material divides them into three general types. In one the surface is regular and horizontal, in another the surface is regular but inclined, and in the third the surface is irregular.

[1] O. E. Meinzer, "Geology and Water Resources of Big Smoky, Clayton, and Alkali Spring Valleys, Nevada," *U.S. Geol. Survey, Water-Supply Paper 423* (1917), pp. 86, 87, Plate II.

[2] *Ibid.*, p. 89.

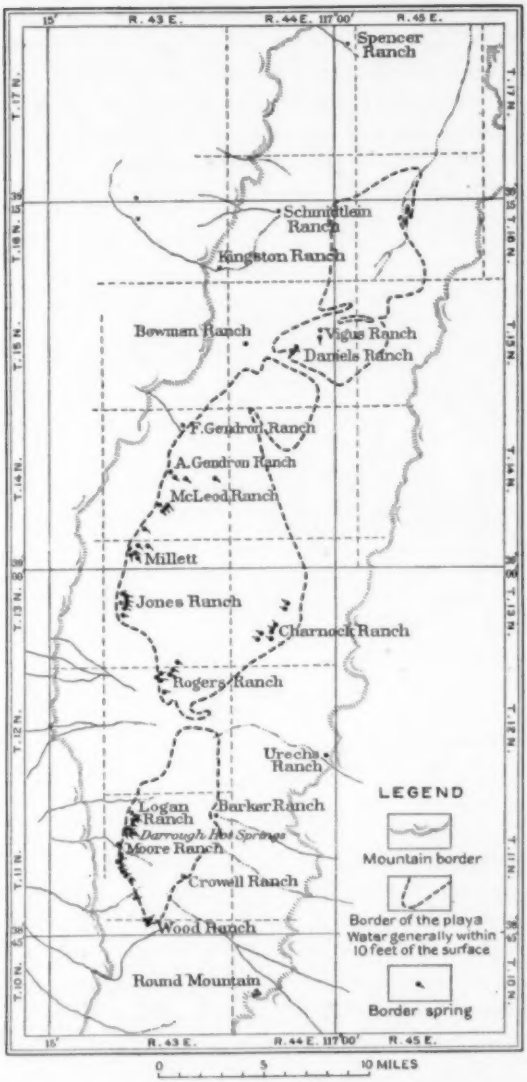


FIG. 5.—Map of the upper Big Smoky Valley, Nevada, showing border springs. (After Meinzer.)

1. *Springs at the outcrop of a horizontal surface.*—Where the impervious rock has a horizontal and regular surface of large extent the rock is usually a member of the sedimentary series of which the overlying porous material is a part. Exceptions, however, occur where the porous material is a surficial deposit but is so regular in its thickness as to give a similar result, or where it is of volcanic origin. There are three classes of these springs.

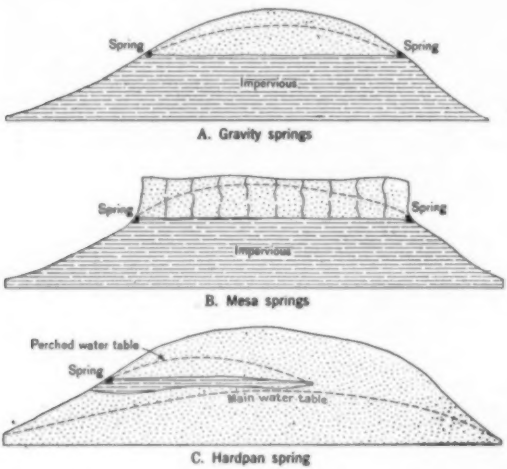


FIG. 6.—Diagram illustrating three classes of contact springs with underlying bed regular and horizontal: *a*, gravity springs; *b*, mesa springs; *c*, hardpan springs.

Gravity springs are those which issue at the contact of a soft and previous bed with an underlying impervious bed (Fig. 6 *a*). The term has been used by Fuller[1] and others. As all springs of the shallow-water type are due to gravity, this term has only the merit of usage. Good examples are Camp Grounds Spring, Crab Tree Spring, and others east of Austin, Arkansas. These springs occur around the edges of the sand-hill area, a ridge of Tertiary rocks on the border of the Mississippi embayment (Fig. 7). Here a thin

[1] M. L. Fuller, "Underground Waters for Farm Use," *U.S. Geol. Survey, Water-Supply Paper 255* (1910), p. 22.

sandstone and sandy clay overlie clay. Water that collects in the sandy beds emerges in a series of springs at the contact with the underlying clay, slightly above the level of the adjacent flood plain of the Mississippi.[1]

Mesa springs are those due to an overlying material which is hard and forms a cliff (Fig. 6 *b*). In this class the overlying material is commonly a sandstone, though it may be a porous and jointed lava flow. Along the east side of Chuska Mountain, New Mexico, in the Navajo country, at an elevation of 8,200 feet, a rough-floored terrace extends for several miles (Fig. 8). Above the shale which floors the terrace a cliff of horizontal Chuska sandstone rises 200 to 500 feet. At the foot of the cliff are more than thirty springs.[2]

FIG. 7.—Cross-section from Austin to Hickory Plains, Arkansas, illustrating gravity springs. (After Purdue.)

If the underlying bed is of small extent but impervious, it will force water contained in overlying porous material to the surface. In many places such water lies far above the ordinary water table and constitutes what is called a perched water table.[3] Such conditions are fairly common in unconsolidated alluvium. The impervious bed is usually clay, cemented gravel, a "mortar bed," or a layer of caliche or hardpan. As in popular usage all these materials are known as hardpan, springs caused by them may be called hardpan springs (Fig. 6 *c*). The older alluvium of the Sacramento Valley is generally underlain by a bed of reddish hardpan about two feet below the surface. During and after rains water seeps to the surface at the contact of the hardpan with the overlying soil, forming temporary springs. A few springs of this class

[1] A. H. Purdue, "Water Resources of the Contact Region between the Paleozoic and Mississippi Embayment Deposits in Northern Arkansas," *U.S. Geol. Survey, Water-Supply Paper 145* (1905), pp. 93, 113.

[2] H. E. Gregory, "The Navajo Country," *U.S. Geol. Survey, Water-Supply Paper 380* (1916), pp. 140–41.

[3] A. C. Veatch, "Underground Water Resources of Long Island, N.Y.," *U.S. Geol. Survey, Prof. Paper 44* (1906), p. 57.

near Corning, California, in places where the soil above the hardpan is thick and continuous enough to form a reservoir, are permanent.

2. *Springs at the outcrop of an inclined surface.*—If the underlying impervious bed has a regular but inclined surface, all the

FIG. 8.—Diagram illustrating the conditions producing springs on the east flank of Chuska Mountain, New Mexico. (From Gregory.)

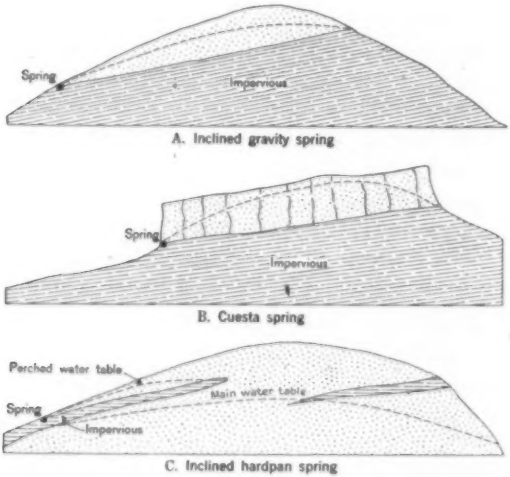


FIG. 9.—Diagram illustrating three classes of contact springs with underlying bed regular and inclined: *a*, inclined gravity springs; *b*, cuesta springs; *c*, inclined hardpan springs.

springs will occur at the outcrop of the contact on the low side, unless the overlying bed is very thick and the dip of the rocks slight. In general, where the underlying bed is of large extent, these conditions exist in sedimentary rocks. Springs of this kind are

divided into two classes, one in which the overlying material is soft and the other in which it is hard. Those of the first class may be called inclined gravity springs (Fig. 9 *a*) and those of the second cuesta springs, after the topographic feature analogous to a mesa (Fig. 9 *b*). Large springs of the inclined gravity type occur near Baden-Baden, Germany (Fig. 10). At this place the lower and middle Buntsandstein, both porous and water-bearing, rest on the smooth erosion surface of the granite. The sandstone has been tilted and eroded into isolated patches. Numerous springs issue along the contact of the sandstone and the granite on the lower side of these erosion remnants. A few springs occur in the talus slopes and on the higher contacts.[1]

Good examples of cuesta springs are those which arise at the contact of basaltic lava and shales of Tertiary age along the western flank of South Table Mountain, north of Oroville, California. The lava is sufficiently jointed and porous to collect water, which runs down the dip of the contact and emerges along the western and lower face of the cuesta.

If the underlying bed is of small extent, springs can only occur where the impervious layer dips with the slope of the hillside but at a smaller angle. These springs are essentially like those in which the impervious layer is horizontal and may also be called hardpan springs (Fig. 9 *c*). The Mountain Mist Springs, in the West Hills, along the northern shore of Long Island (Fig. 11), are of this class. The underlying hardpan is a compact till, and the perched water table lies about 200 feet above the main water table.[2]

3. *Springs at the outcrop of an irregular surface.*—Where the underlying impervious bed has an irregular surface, the overlying material is commonly thin and unconsolidated—a mantle overlying the bed rock. In humid countries the great majority of springs are formed by water that collects in the reservoirs afforded by this porous material. It varies in thickness and character and the irregularities of the underlying surface originate in diverse ways.

[1] H. Eck, "Geognostische Beschreibung der Gegend von Baden-Baden, Rothenfels, Gernsbach, und Herrenalb," *K.-preuss. geol. Landesanstalt Abh.*, neue Folge, Heft 6 (1892), pp. 653 ff.

[2] A. C. Veatch, "Underground Water Resources of Long Island, N.Y.," *U.S. Geol. Survey, Prof. Paper 44* (1906), pp. 57–58.

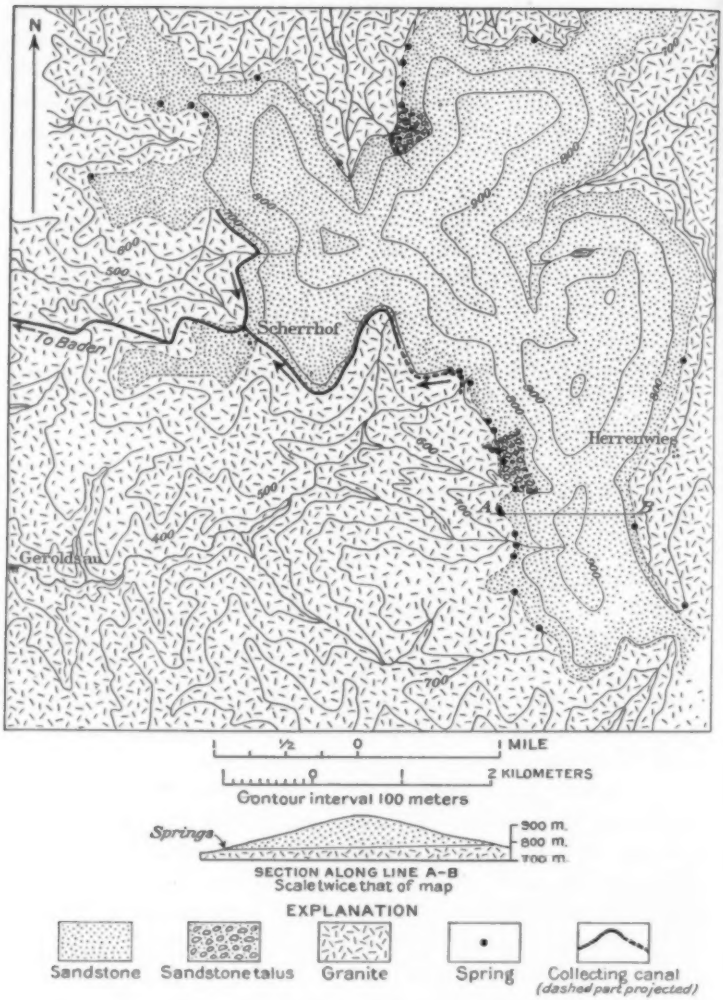


FIG. 10.—Map of an area near Baden-Baden, Germany, showing inclined gravity springs, and connecting canal fed by inclined gravity springs. (Redrawn from Keilhack, after Eck and Lueger.)

Four classes of springs in this group are distinguished in the following paragraphs.

Where the overlying material is thick and of wide extent, the contact is ordinarily an unconformity. The irregularities of such a contact are of minor significance, and here all the types of springs included in the previous group—gravity, inclined gravity, cuesta, and mesa springs—may occur. The springs will be found at the

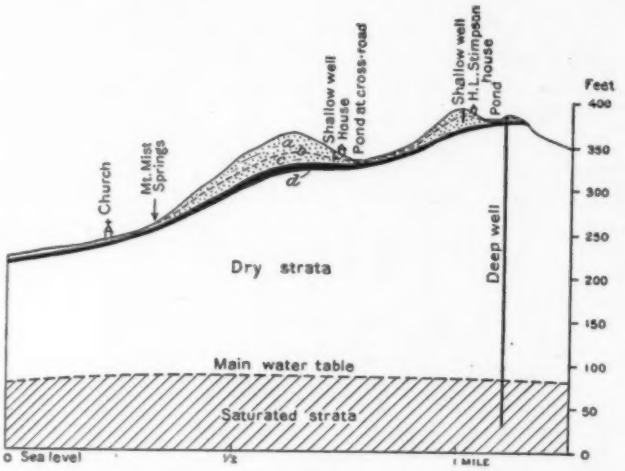


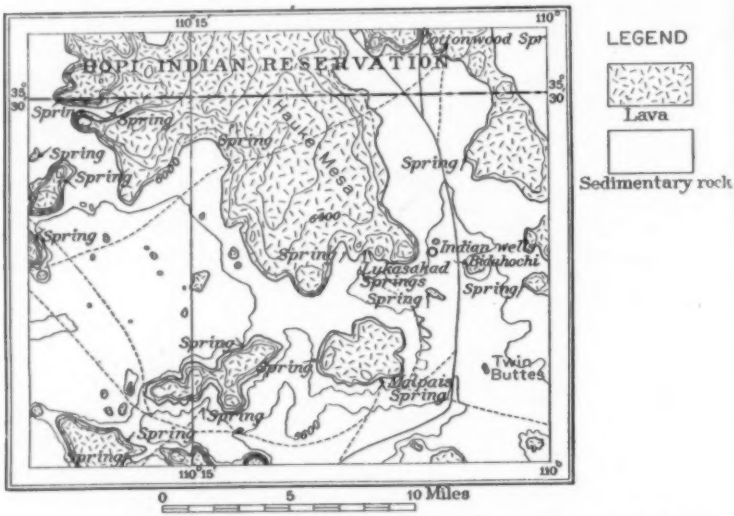
FIG. 11.—Cross-section showing inclined hardpan on the north side of the West Hills, Long Island, New York, and source of Mountain Mist Springs. (After Veatch.)

lowest parts of the contact. Mesa springs that fulfil these conditions occur at the Hopi Buttes, in Arizona (Fig. 12), where the overlying bed is a porous and jointed lava flow underlain by patches of volcanic ash and the underlying impervious beds are tilted and eroded shales and sandstones. The unconformity is fairly smooth, but springs issue in the lowest places.[1]

Where the overlying porous material is localized and the contact is irregular, the resulting springs may be called pocket springs, because the reservoir from which the water is drawn lies

[1] H. E. Gregory, *op. cit.*, p. 139.

in a pocket in the underlying rock (Fig. 13 *a*). The overlying material, which is usually unconsolidated and generally of relatively small volume, may be of different kinds, of which the most common are residual soil, alluvium, till, and wind-blown sand. This is the prevailing type of spring in the till-covered hills of New England and in the soil-covered crystalline rocks of the Sierra Nevada. Most small springs are grouped under this head.


110°15'
110°
HOPI INDIAN RESERVATION
Cottonwood Spr
35°30'
Spring
Spring
Spring

Overflow springs, like pocket springs, are due to the overflow of a reservoir of porous rock (Fig. 13 *b*). The underlying impervious bed is not continuous, but percolation through the porous bed is not sufficient to drain the reservoir. The reservoir is commonly large and of structural origin. These springs are most common in

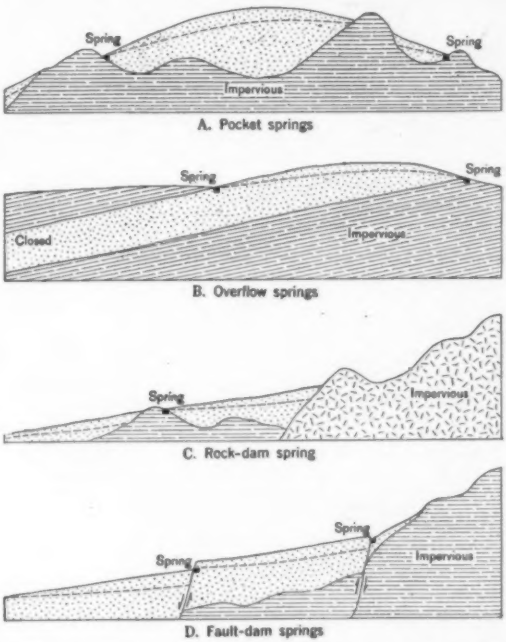


FIG. 13.—Diagram illustrating four classes of contact springs with underlying bed irregular: *a*, pocket springs; *b*, overflow springs; *c*, rock-dam springs; *d*, fault-dam springs.

the collecting area of an artesian system and are distributed along the contacts of the inclosing impervious rocks. An example is found in the North Downs, near London, England (Fig. 14). Many of the springs at this place have ceased to flow because of excessive pumping from wells in the artesian basin to the south.[1]

[1] H. B. Woodward, *The Geology of Water Supply* (London, 1919), pp. 133–35, Figs. 12, 34.

Similar masses of pervious rock may be bounded on one or more sides by impervious rock brought into place by faulting. Overflow will then take place among the fault line, as is exemplified in the northern Vosges Mountains, France[1] (Fig. 15).

In arid regions, particularly, great thicknesses of alluvium have accumulated on irregular rock floors, and if the irregularities of

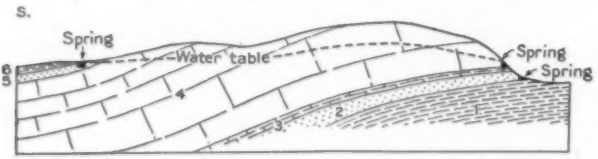


FIG. 14.—Cross-section of the North Downs, near London, England: 1, gault; 2, upper greensand; 3, chalk marl; 4, main chalk, pervious to water; 5, lower London Tertiary strata; 6, London clay, impervious to water. (Redrawn from Woodward.)

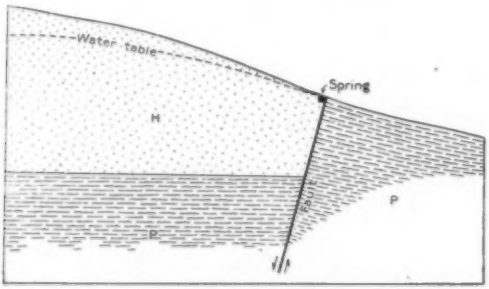


FIG. 15.—Cross-section showing overflow spring in the northern Vosges Mountains, France. H. Hauptbunt-sandstein (porous); P. Rötelschiefer (impervious). (After Leppla.)

the rock floor project through the surface of the alluvial plains, the water contained in the alluvium may be forced to the surface. Essentially similar conditions may be brought about by intrusion of igneous material into such alluvial plains. Under these conditions the rock projections act as dams against the regular flow of the ground water, and the resulting springs may be called rock

[1] A. Leppla, quoted by K. Keilhack, *Lehrbuch der Grundwasser und Quellenkunde*, Berlin, 1912.

dam springs (Fig. 13 *c*). The best-known examples are the ciénegas of the valleys of southern California. West of Cucamonga (Fig. 16) a projection of older cemented alluvium in the Red Hills

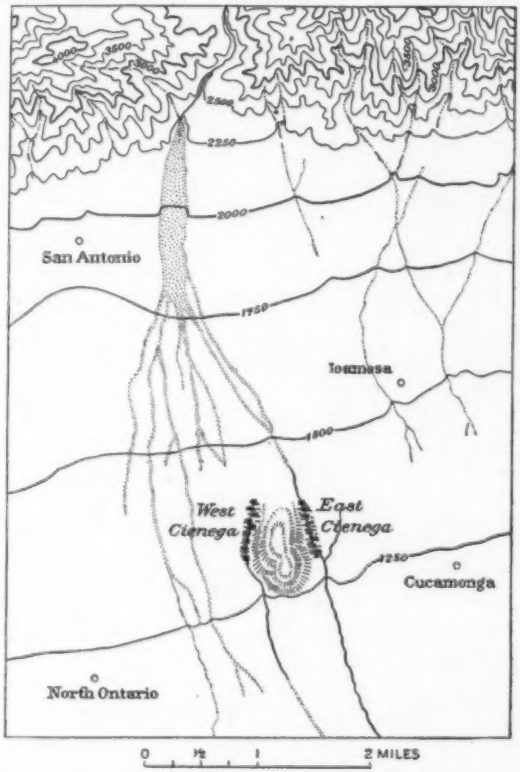


FIG. 16.—Map showing rock-dam springs near Cucamonga, California. (After Mendenhall.)

produces East and West ciénegas; a similar projection of the rocks of the San Rafael Hills produces the springs of Devil's Gate, north of Pasadena.[1] In the Antelope Valley a part of the rock floor of

[1] W. C. Mendenhall, "Ground Water and Irrigation Enterprises in the Foothill Belt, Southern California," *U.S. Geol. Survey, Water-Supply Paper 219* (1908), pp. 34–38, 50–53.

the basin rises above the alluvial plain and produces Lovejoy Springs.[1] Examples of rock dams formed by intrusion of igneous rock into alluvium have not yet been found.

Fault dam springs arise where the free flow of ground water through porous material is interfered with by faulting (Fig. 13 *d*). Fault zones become efficient barriers by the formation of gouge or simply through disturbance and dislocation of the beds. Examples

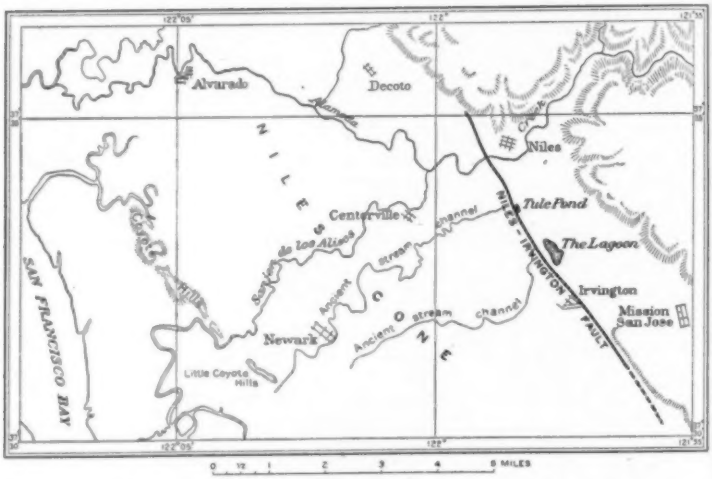


FIG. 17.—Map of the Niles cone, California, showing fault-dam springs due to the Niles-Irvington fault. (After Clark.)

of this class of spring have been found in the Big Smoky Valley, Nevada, by Meinzer.[2] The Niles-Irvington fault, in the Santa Clara Valley, California, beheads an alluvial fan (the Niles cone[3]), as shown in Figure 17. On the side near the mountains the water

[1] H. R. Johnson, "Water Resources of Antelope Valley, California," *U.S. Geol. Survey, Water-Supply Paper 278* (1911), p. 52, Fig. 9.

[2] O. E. Meinzer, "Geology and Water Resources of Big Smoky, Clayton, and Alkalai Spring Valleys, Nevada," *U.S. Geol. Survey, Water-Supply Paper 423* (1917), p. 90.

[3] W. O. Clarke, "Ground Water Resources of the Niles Cone and Adjacent Areas, California," *U.S. Geol. Survey, Water-Supply Paper 345* (1915), pp. 130–32, 150.

table is about 20 feet higher than on the downstream side. Two ponds fed by ground water lie east of the fault. They are fault dam springs.

C. SPRINGS IN POROUS ROCK BETWEEN IMPERVIOUS ROCK (ARTESIAN SPRINGS)

Where water is contained in the pore spaces of a pervious bed lying between impervious strata, the essential conditions for the

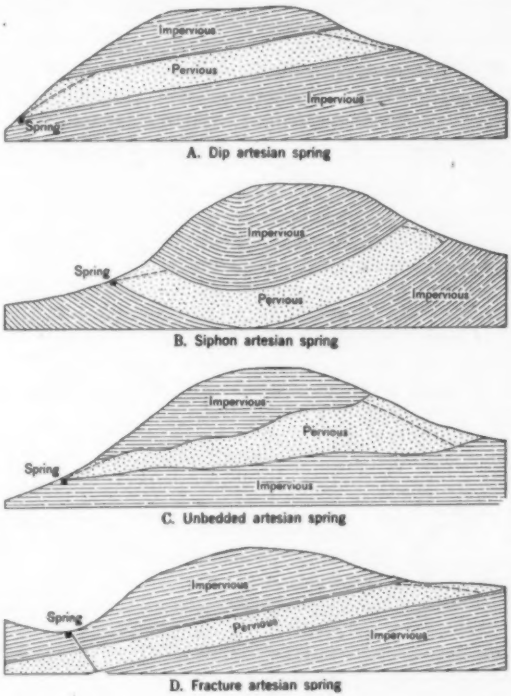


FIG. 18.—Diagrams illustrating four classes of artesian springs: *a*, dip artesian springs; *b*, siphon artesian springs; *c*, unbedded artesian springs; *d*, fracture artesian springs.

existence of springs are that a part of the porous bed outcrop so as to absorb rain in order to provide a regular supply of water, and that the beds be inclined—in other words, the essential conditions

are those of an artesian basin. There are four classes of springs which fulfil these conditions.

Dip artesian springs (Fig. 18 *a*) occur in more or less regularly bedded rocks that have been tilted and eroded in such a manner that the porous bed receives water from the rain or streams in its upper end and that the lower end is exposed at the surface. Sedimentary rocks, alternating lava flows, tuffs and gravels, and unconsolidated alluvial material supply these conditions.

In a series of beds that has been folded, if the proper conditions of inflow and outflow exist, a porous stratum constitutes an inverted siphon for the conveyance of water. Springs due to the outflow from the low side of such a basin may be called siphon artesian springs (Fig. 18 *b*). The artesian basin of the Great Plains is the

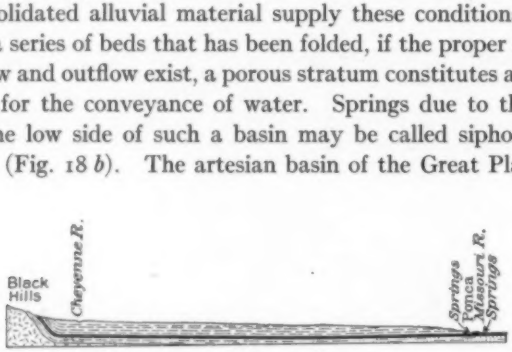


FIG. 19.—Cross-section of the Great Plains showing siphon artesian springs near Sioux City, Iowa. (After Darton.)

largest and most remarkable in the world, because of the great distance traveled by the water and the heavy pressure under which it exists. Along the eastern border of the basin the beds have a distinct westward tilt, but the water-bearing Dakota sandstone is not everywhere exposed (Fig. 19). However, near Sioux City, Iowa, numerous springs and seeps are found along the western bluffs of Missouri River, where the Dakota first emerges from beneath its impervious cover.[1]

In certain unconsolidated deposits not regularly bedded, a mass of porous material may be so exposed as to receive water at a high level and discharge this water at a lower level. The springs resulting from this structure may be called unbedded artesian

[1] G. E. Condra, "Geology and Water Resources of a Portion of the Missouri River Valley in Northeastern Nebraska," *U.S. Geol. Survey, Water-Supply Paper 215* (1908), pp. 27, 28.

springs (Fig. 18 *c*). Though rare, springs of this class occur in till, and doubtless examples may be found in other types of rock.

The springs of the three classes just described depend on the outcrop of the saturated porous bed in its lower portion. Springs occurring where the porous bed does not crop out, but the water escapes from it by an opening leading to the surface, may be termed fracture artesian springs (Fig. 18 *d*). Many of these springs have been classed with fissure springs, but it seems essential to distinguish between springs that tap artesian basins and yield water under hydrostatic head, and those that depend on the deeper waters of the crust. This name is chosen because it seems probable that all such openings are primarily fractures in the rocks. Although it is true that great pressure is exerted by the water in deep artesian basins, it seems improbable that pressure could force water to rise through overlying material in sufficiently definite channels to supply springs unless it moved along pre-existing faults or other fractures. Water pressure may assist in keeping the fracture open, and the flow of water tends to plaster up caving walls with mud, as in the hydraulic systems of drilling wells, or to solidify them with precipitated minerals. Many of the springs of this class have a high temperature, a steady flow, and an alignment that causes them to be confused with true fault springs. However, a distinction should be made. If definite artesian structure can be demonstrated, most of the water is probably meteoric in origin—that is, it is derived from rain or melted snow absorbed at the outcrop; if artesian structure cannot be demonstrated, there is a strong presumption that much of the water arising along great faults and deep fractures in the crust is of juvenile origin—that is, it issues from deep within the earth and now sees the light of day for the first time. At any rate, there is a sufficiently distinct difference in structure to justify separate classification of the springs.

In Antelope Valley, California, porous sand and gravel underlying clay contain water under artesian pressure. Numerous wells have been sunk to obtain this water, and its chemical character and temperature are known. Buckhorn, Indian, Willow, and other springs having water similar in character and temperature

lie along a fault at the foot of Rosamond Buttes but probably not exactly on the fault trace (Fig. 20). This fault is represented by an escarpment in the alluvium west of Willow Springs, from 50 to 100 feet in height and extending for a distance of 5 miles.[1] The

FIG. 20.—Map of Antelope Valley, California, showing fracture artesian springs. (After Johnson.)

irregular alignment of these springs is best explained by assuming that there are associated with the main fault a number of subsidiary but parallel fractures which permit the artesian water to rise. Alternative explanations are offered by Johnson, but the absence of springs except along this general line makes other

[1] H. R. Johnson, *op. cit.*, pp. 21 and 49.

explanations improbable. These springs, therefore, may be taken as typical of the class of fracture artesian springs.

D. SPRINGS IN IMPERVIOUS ROCK

The term porous and impervious are of course relative, but impervious rocks may be considered to have pore spaces of capillary or smaller size, which usually form a relatively small percentage of the rock. Some clay, however, has a high porosity but because of the absorption of water and swelling of individual grains is highly impervious. Through such pore spaces effective flow under hydrostatic head is impossible. Movement takes place

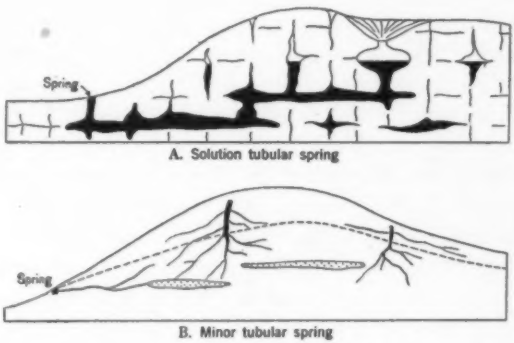


FIG. 21.—Diagram illustrating two classes of tubular springs: *a*, solution tubular springs; *b*, minor tubular springs.

through openings which may be grouped into two types: (1) those which are more or less tubular in form, and (2) those which are sheetlike in form and along which the water moves as a thin film.

1. *Tubular Springs.*—Springs of the first group may be called tubular springs. This group can in turn be divided into three classes, according to the origin of the tubular openings. Solution tubular or cavern springs are due to channels and openings formed through solution of the rock by circulating ground water (Fig. 21 *a*). Solution commonly begins along joints or other previously existing openings, and the channels may be enlarged to a very great size. In these tubes the water then flows freely and with relative rapidity.

The rocks that are affected by this process are limestone, calcareous sandstone, gypsum, and salt. Silver Spring, in Marion County, Florida, is a well-known example. This spring emerges from several openings in the cavernous Vicksburg limestone into a pool several acres in extent. The volume of flow is about 385,000 gallons a minute, or 855 cubic feet a second, a stream sufficiently large to float a steamer.[1]

Lava tubular springs are due to caverns and tunnels in lava flows. These caverns and tunnels are formed through the process of igneous extrusion. When a very liquid lava cools rapidly, a crust of sufficient strength to support itself may form on the surface and the liquid lava below may flow out, leaving an arched tube. Under other conditions the crust may yield to lateral pressure of the underlying liquid lava and form arched tunnels which are made permanent by the solidification of the lava. Principally by these two processes caverns and tunnels may be produced in formations due to successive lava flows, and under favorable circumstances such openings may form channels for the ground water. On the southern flank of the mountain mass which culminates in Lassen Peak, California, large springs break out in Big Meadows, about 5 miles from Prattville.[2] The water issues from basaltic lava in an area about 100 yards in diameter. The flow is 29,000 gallons a minute, or 62.6 cubic feet a second, and the temperature is 46° F. The springs belong to a series characteristic of the lower slopes of the Lassen Peak mass, and a few of them are shown in Figure 2 (p. 533). The low temperature of these springs indicates that the water originates high up on the mountain and flows rather quickly down the slope. It seems probable that the water flows through caverns and tunnels in the lava rather than through the pores of the rock. This interpretation is supported by the existence of a similar series of springs around Mount Shasta. Plutos Cave, north of this mountain, is an example of the kind of caverns through which the water passes. It is in places 60 feet high and 50 feet wide and

[1] G. C. Matson and Samuel Sanford, "Geology and Ground Waters of Florida," *U.S. Geol. Survey, Water-Supply Paper 319* (1913), p. 367.

[2] G. A. Waring, "Springs of California," *U.S. Geol. Survey, Water-Supply Paper 338* (1915), p. 330.

has been explored for more than a mile.[1] Certain cold springs on Mount Shasta, however, appear to be due to the melting of buried ice.[2]

Minor tubular springs are due to a variety of causes (Fig. 21 *b*). The tubes or channels are of small size and irregular length, and many of them appear to be due, at least in part, to the movement

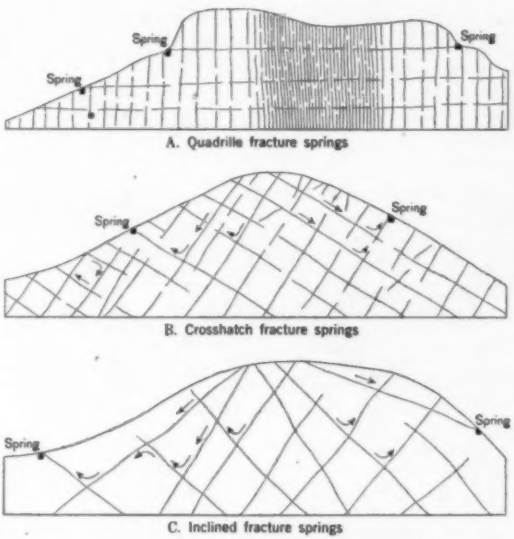


FIG. 22.—Diagram illustrating three classes of fracture springs: *a*, quadrille fracture springs; *b*, crosshatch fracture springs; *c*, inclined fracture springs.

of the water itself. Unconsolidated deposits are particularly susceptible to this action. The decay of plant roots, the existence of small sand streaks, and the enlargements of shrinkage cracks may all assist in the production of these minor openings.

2. *Fracture springs.*—Springs that issue from sheetlike or platelike openings in non-porous rocks may be called fracture springs, because such openings are primarily due to fractures. Joints,

[1] H. W. Fairbanks, *Practical Physiography* (Boston, 1906), p. 178, Fig. 180.

[2] G. A. Waring, *op. cit.*, p. 332.

bedding planes, faults, columnar joints, and openings formed on slaty cleavage, fissility, and schistosity are the common types of fractures in igneous and metamorphic rocks. In sedimentary rocks joints, bedding planes, cross-bedding planes, and faults are the principal types of fractures. If the fractures are very numerous and closely spaced, however, water moves through the rock in essentially the same fashion as in porous rocks. Minute fracture systems add very much to the effectiveness of interstitial pores and are probably present in all porous rocks. The essential difference is that in a porous rock water moves bodily, and, as a rule, slowly

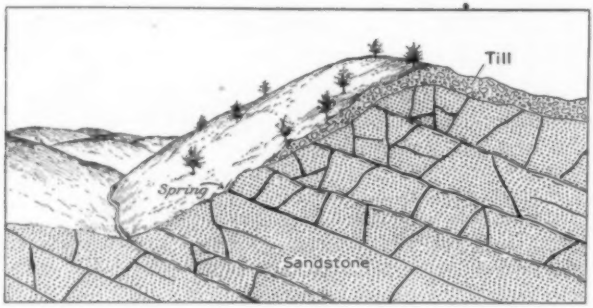


FIG. 23.—Diagram showing occurrence of springs near Mount Carmel, Connecticut. (After Gregory.)

through minute openings, whereas in fractured impervious rocks it moves more rapidly as sheets, threads, or films of water.

Fracture springs may be divided into two classes according to the attitude of the fractures toward one another and toward the horizon. A common joint system is one in which there are two or three sets of joints at right angles to one another and one of the sets is horizontal. Springs due to such joints may be called quadrille fracture springs (Fig. 22 *a*). Such a jointing system is common in sedimentary rocks and in impervious igneous rocks, particularly in sheeted plutonic rocks. Springs due to two systems of joints at right angles to each other and inclined to the horizon may be called crosshatch fracture springs (Fig. 22 *b*). Joints and bedding planes in Triassic sandstones form such a system of

fractures and give rise to springs near Mount Carmel, Connecticut, as shown in Figure 23.[1] Far more abundant, however, are irregular systems of joints, all of which are inclined toward the horizon, and springs due to such systems may be called inclined fracture springs (Fig. 22 *c*).

KEY TO THE CLASSIFICATION OF SPRINGS

I. Springs due to deep-seated waters, juvenile and connate, admixed with deeper meteoric water; do not flow under hydrostatic head and are usually not subject to seasonal fluctuation.

- A. *Volcanic Springs.* Associated with volcanism or volcanic rocks; water commonly hot, highly mineralized and containing gases. Grade from gas vents into springs of normal temperature indistinguishable from those due to other causes.
- B. *Fissure Springs.* Due to fractures extending into deeper parts of the crust; water usually highly mineralized and commonly warm or hot.
  1. Fault Springs. Associated with recent faults of great magnitude.
  2. Fissure Springs. No direct structural evidence as to origin, but because of temperature and steady flow believed to have deep origin.

II. Springs due to meteoric and occasionally other waters moving as ground water under hydrostatic head; many fluctuate in flow with the rainfall.

- A. *Depression Springs.* Due to land surface cutting water table in porous rocks.
  1. Dimple Springs. Due to depressions in hillsides.
  2. Valley Springs. Due to abrupt change in slope at edge of flood plain.
  3. Channel Springs. Due to depressions in flood plains or alluvial plains caused by channel cutting of stream.
  4. Border Springs. Due to change in slope at border between alluvial plains and playas, lake beds, or river bottoms; relative imperviousness of central clay deposits assists flow.
- B. *Contact Springs.* Due to porous rock overlying impervious rock.
  1. Impervious rock has a horizontal and regular surface.
     a) Underlying bed is of large extent; common in consolidated sedimentary rock.
        (1) Gravity Springs. Overlying material is soft.
        (2) Mesa Springs. The overlying material is hard, usually sandstone or lava flow; water contained in pores and joints of the rock.

[1] H. E. Gregory and E. E. Ellis, "Underground Water Resources of Connecticut," *U.S. Geol. Survey, Water-Supply Paper 232* (1909), p. 136.

b) Underlying bed is of small extent; common in unconsolidated alluvium; impervious bed is usually clay, cemented gravel, "mortar bed," caliche, or hardpan.

(1) Hardpan Springs.

2. Impervious bed has an inclined and regular surface; all springs on the low side unless the overlying bed is very thick and the dip low.

a) Underlying bed is of large extent.

(1) Inclined Gravity Springs. The overlying material is soft.

(2) Cuesta Springs. The overlying material is hard; of same character as mesa springs.

b) Underlying bed is of small extent; as in hardpan springs.

(1) Impervious layer dips away from hill; spring possible.

(2) Impervious layer dips into hill; spring possible only in ravines.

3. Impervious bed has irregular surface.

a) Overlying porous material is thick and of wide extent; contact is unconformity. Gravity, inclined gravity, mesa, and cuesta springs may occur, but springs will be sharply localized at lowest parts of contact.

b) Pocket Springs. Overlying porous material is unconsolidated and more or less discontinuous, residual soil, talus, landslide débris, alluvium, till, stratified drift, wind-blown sand, or volcanic ash.

c) Overflow Springs. Irregular floor is not continuous, but porous bed is saturated and overflows at lateral contacts; common at receiving end of artesian systems.

d) Rock Dam Springs. Irregularities of the rock floor under an alluvial plain force water to surface; these may be projections of floor of basin, projections of partly consolidated older alluvium, igneous dikes, or volcanic plugs.

e) Fault Dam Springs. Dam caused by faulting.

C. *Artesian Springs.* Due to pervious bed between impervious materials.

1. Dip Artesian Springs. More or less regularly bedded rocks; tilted porous bed crops out in valley; usually sedimentary, also alternations of lava flows, flow breccias, tuffs, gravels.

2. Siphon Artesian Springs. Similar rocks; folded and with outcrops in valley.

3. Unbedded Artesian Springs. Rocks not regularly bedded, but mass of porous material is exposed so as to receive water and crops out in valley; occur in till and perhaps in other rocks.

4. Fracture Artesian Springs. All the conditions above, except that lower end of porous bed does not crop out but an opening allows water to escape. Opening due to fracturing with or without faulting.

D. *Springs in Impervious Rock:*

1. Tubular Springs. Due to more or less rounded channels in impervious rocks.
   - *a*) Solution Tubular or Cavern Springs. Due to solution channels in limestones, calcareous sandstones, gypsum, salt.
   - *b*) Lava Tubular Springs. Due to caverns and tunnels in lava flows.
   - *c*) Minor Tubular Springs. Due to channels made by movement of water, decay of tree roots, sand streaks, or shrinkage cracks, usually in unconsolidated sediments.

2. Fracture Springs. Due to fractures consisting of joints, bedding planes, columnar joints, openings due to cleavage, fissility, schistosity, cross-bedding planes, and faults in impervious sedimentary, igneous, and metamorphic rocks.
   - *a*) Quadrille Fracture Springs. Due to more or less rectangular system of fractures, one of which is parallel to the horizon.
   - *b*) Crosshatch Fracture Springs. Due to more or less rectangular system of fractures, inclined toward the horizon.
   - *c*) Inclined Fracture Springs. Due to inclined fractures, not necessarily systematic.

AEQUINOCTIA, AN OLD PALEOZOIC CONTINENT

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Wassenaar, Holland

Through the whole of Central Celebes,[1] from the Gulf of Boni to the Gulf of Tomini, extends a formation of crystalline schists, lying east, but principally west, of Lake Posso. These schists are strongly folded and their positions vary from horizontal to vertical. More than once, but last during the Oligocene, this region was leveled by denudation. The resulting peneplain was arched during the neo-Tertiary and Plio-Pleistocene to 2,000 meters above sea-level. Consequently, it became deeply cut by erosion. In some of these furrows one can distinctly observe an east-west strike of the strata, which seems to be the effect of the oldest folding. Especially on the west of Lake Posso the region appears now as a distinct geographical entity. West of these mounts others follow, either separated from the first by depressions or not. They consist of gneiss, viz., biotite and amphibole gneiss, amphibolites, granites, and effusive rocks. In this part of the country the peneplain character has not been preserved so well; the highest tops reach altitudes up to 3,000 meters.

In no other part of the Dutch East Indian Archipelago do the gneiss and schists region occupy such a large surface as in Central Celebes; besides that, the crystalline schists are characterized by a particular richness of rock varieties.

Most of the schists belong to the second and eighth groups of the Grubenmann[2] classification, and to the upper and middle zones of these groups. This leads to the conclusion that they represent metamorphosed clays, sandstones, and argillaceous sandstones. Besides these rocks, a certain importance must be attributed to the metamorphosed calcareous sandstones and limestones, rocks of the

[1] See E. C. Abendanon, *Voyages géologiques et géographiques à travers la Célèbes Centrale*, 3 volumes and one atlas, Leyde, Holland, 1916–18.

[2] U. Grubenmann, *Die kristallinen Schiefer*, Vol. II, 1907.

ninth and tenth groups, some of them making part of the upper zone, but chiefly belonging to the middle zone. These latter varieties appear principally along the eastern bank of Lake Posso and north of the lake.

With these schists of sedimentary origin, there are others which must have evolved from more or less basic eruptive rocks and their tuffs. But for a few exceptions, belonging to the middle zone, most of these latter schists belong to the upper zone; they are represented by rocks of groups three (plagioclase gneiss), four (metamorphosed gabbroid rocks), and five (schists containing silicates of magnesia).

The age of this formation of schists in Central Celebes and in other parts of the archipelago has been interpreted as follows:

1. Wichmann[1] considers the mica schists of Central Celebes as Archean; Martin[2] does the same regarding those of Ceram, whereas Verbeek[3] thinks that those of South Borneo constitute a series of Azoic rocks; Volz[4] refers to the Archean only the gneiss of the land of the Gajos (North Sumatra).

2. Molengraaff[5] believes that the crystalline schists of Central Borneo are partly Archean, partly of more recent age.

3. Verbeek[6] groups the gneiss and mica schists which he found in the Moluccas, with other rocks, such as the quartzites (of dark color) and argillaceous schists, in the "old schists formation," in which he presumes the existence of Azoic (e.g., the gneiss of Sibelaboe) or Archean elements, but also elements of the Old Paleozoic. His opinion upon this matter is shared by Volz[7] and Ahlburg,[8] who also connect other rocks, such as the peridotite of eastern and southeastern Celebes, with the mica schists. Volz draws attention to the complete conformability between the mica schists and the old schists in the aforementioned land of the Gajos. It must be

[1] "Der Posso-See in Celebes," in Petermann's *Geogr. Milteilungen*, Heft VII, 1896.

[2] "Reisen in den Molukken," *Geologischer Theil*, 1903.

[3] *Jaarboek van het Mijnwezen*, Vol. I, 1875.

[4] *Nord-Sumatra*, Vol. II, 1912.

[5] *Geologische verkenningstochten in Centraal. Borneo*, 1900.

[6] *Rapport sur les Moluques*, 1908.

[7] *Nord-Sumatra*, Vol. I, 1909.

[8] *Versuch einer geologischen Darstellung der Insel Celebes*, 1913.

remarked that a similar conformability has been discovered nowhere else in the archipelago by any of the other explorers.

4. The rare true mica schists of Letti are regarded by Molengraaff[1] as metamorphosed rocks of the Old Permian.

5. By Wing Easton,[2] but more expressly by Volz and Tobler,[3] the "old" schists are considered as being partially Mesozoic. One may doubt, however, whether they include real mica schists.

6. Smith[4] expresses much doubt about the Archean age ascribed to the crystalline schists of the Philippine Islands, and he assigns many of them to the Tertiary.

There is thus a strong discrepancy in the opinions of these competent explorers, as the ages they assign to the schists vary from the Archean to the Tertiary. Their opinions could hardly differ more.

In order to come to a personal conclusion, I shall begin by trying to solve a few questions. First of all: Do all the series of rocks which the above-mentioned explorers have classed with the "old schists formation," really belong to it?

It would perhaps be possible to answer this question with certainty if one had gathered into a single collection all the rocks collected; but such a collection does not exist. On the contrary, those assembled by the explorers named are scattered in such a large number of museums that a comparative petrographic examination of the crystalline schists and old schists existing in the different parts of the Dutch East Indian Archipelago is absolutely impossible. One thing is certain, however, and that is that the rocks included in the "old schists formation" have been put together on a basis of negative characters, namely, their "old" appearance and the absence of fossils. It is therefore possible, perhaps very probable, that elements have been included which in fact do not belong to the old schists formation.

In order to come to a more accurate classification, it appears necessary to make, first of all, a distinction between the gneiss and crystalline schists on the one side, and the "old" schists on the other

[1] *Jaarb. Mijnwezen*, 1914. [2] *Ibid.*, 1904.

[3] *Voorl. med. over de geologie der res. Djambi*, 1912.

[4] *Handbuch der regionalem Geologie*, Philippine Islands, 1910.

side. The real gneiss must be separated from the gneissoid rocks which originated at the periphery of the younger granites at the time of their solidification. These gneissoid rocks appear in large quantity in Central Celebes, and Volz believes he has detected them in the north of Sumatra. As to the crystalline schists, it is advisable to distinguish the old mica schists from the rocks belonging to the more recent contact zones, as for instance the andalusite mica schist from Central Celebes. Finally, there is no doubt that there has been included with the "old" schists rocks[1] which do not belong to them. For instance, in his *Rapport sur les Moluques* (p. 755), Verbeek places the argillaceous schists of the island of Taliabo in the old schists formation. However, they recall so much certain schists of the Upper Cretaceous, either metamorphosed or not by the granite of the southwestern part of Central Celebes, that I can no longer consider them as being undoubtedly old schists.

Having eliminated the rocks which do not belong to them, it must be accepted that there exists in the Dutch East Indian Archipelago three series of rocks which are unquestionably of similar age, and which form the base of the other geological formations. Of these, the gneiss and mica schists are the earliest member. This double series appears as large complexes, principally in the north of Sumatra, the southeast of Borneo, on very extensive areas in Central Celebes, and in the islands of Boeroe and Ceram. As in other places, it is possible to see the lowest strata, viz., the gneiss formation, in this latter island and elsewhere in the archipelago, but especially and very clearly in Central Celebes. In its upper part, these crystalline schists seem to grade into the "old" schists, which constitute the second essential member of the oldest formations. These schists are well known in many parts of the archipelago, but only Volz has pointed out their conformability with the mica schists in the north of Sumatra.

Particularly in Central Celebes the mica schists occupy such an extensive surface, vertically as well as horizontally, that there is

[1] Tobler (and before him Wing Easton and Volz) had already called attention to this point; however, he does not exclude them systematically from the denomination "old schists."

no question of explaining them by contact metamorphism (although it is certain that such metamorphism has taken place locally). We have here to deal with regional metamorphism of ancient date. The total thickness of these schists cannot be estimated, even approximately, but it is certainly some thousands of meters. Such a thickness could not have resulted, in Celebes or in other islands of the archipelago, from the sedimentary beds of the Tertiary or of the Mesozoic. So we reach the conclusion that these rocks are pre-Mesozoic.

What is the tectonic relation between the crystalline schists of Central Celebes and those of other parts of the archipelago?

Suess[1] has remarked that Martin observed a large massif of Archean rocks in the islands Boeroe and Ceram; it consists principally of mica schists having an east-west strike. Similar rocks[2] having the same strike have been found, not only in the group of the Péling-Misool islands,[3] but also in the northwestern peninsula of New Guinea, whereas the island of Roon (east of this peninsula) is composed of gneiss. These data have led him to the hypothesis[4] that the islands Ceram and Boeroe are but the prolongation of the central mountain ranges of New Guinea, whereas the group of the Péling-Misool Islands and the northwestern peninsula of New Guinea represent the foreland of the folded land situated farther south. Suess makes of this foreland an old "massif" (Vol. III, Part 3, p. 1036). He would make a geological entity of the territory extending from Celebes to the little island Rossel, southeast of the southeastern point of New Guinea, but he writes (Vol. III, Part 3, p. 1035): "L'énorme distance qui sépare Célèbes de l'île Rossel (Louisiade), ainsi que les grandes lacunes dans nos connaissances, ne permettent pas d'énoncer ici des résultats certains." Further (Vol. III, Part 1, p. 343): "On ne peut pas dire de quelle façon la chaîne des îles orientées de l'E. à l'W. (Obi, Taliabo, etc.) se prolonge à Célèbes."

[1] *La Face de la Terre*, Vol. III, Part 1, p. 317.

[2] *Ibid.*, Vol. III, Part 3, p. 1033.

[3] *Ibid.*, Vol. III, Part 3, p. 1036: "Des terrains anciens apparaissent dans ces îles (de Taliabo à Misool), surmontés de couches mésozoiques non plissées."

[4] *La Face de la Terre*, Vol. III, Part 1, p. 318 and Vol. III, p. 1035.

As a result of the investigation made in Central Celebes, this last point is now elucidated, for the east-west strike is repeated in a very remarkable way in the schists and gneiss formations of this territory, although the older ranges of Central Celebes and Ceram have a northwest to southeast direction, owing to the post-Lutecian folding, whereas younger horsts of Central Celebes have a north-south direction, the result of more recent tectonic events. Moreover, the old massifs of Central Celebes are separated from those of the islands situated to the east of Celebes by a massif of peridotite. This latter rock has great importance in these islands.

Until the contrary is proved, the answer which one may give to the second question indicates thus that the gneiss and schists massifs, from Central Celebes to the island of Roon, are parts or horsts of an old massif which stretched formerly throughout this whole extent. Without being able to assert it dogmatically, one may say that this old massif, folded in an east-west direction, is prolonged to the west, and that it reaches the middle of Borneo, where Molengraaff and other explorers have also found tectonic entities striking almost east and west.[1]

The answers given to the first two questions are not sufficient to define precisely the age of the mica schists. This leads me to look outside the archipelago and to ask the following questions: What is the distribution of the oldest formations outside the limits of the archipelago? In what places do they reveal their age with certainty by the presence of fossils? What geo-metamorphic processes have the formations in those places undergone, and what is the relation of these processes to those which have predominated in the archipelago?

If the metamorphosing processes in other places have not been more favorable than in the archipelago for the preservation of fossils of Paleozoic, and especially of early Paleozoic age,[2] we have one reason less for presuming that fossils once existed in the schists of the archipelago.

[1] The same direction is repeated in the northern peninsula of Celebes, in the isle of Java, and in the chain of the small Sunda Islands.

[2] Neo-Carboniferous and Permian fossils have been found in the archipelago; this is a well-known fact.

To answer the foregoing questions I cannot do better than again to refer to Suess's *La Face de la Terre*.

In the eastern part of New Guinea there exist old rocks much folded and strongly upheaved,[1] which occupy the highest parts of the interior, as for instance the Owen-Stanley Range, Adolf-Haven, and the southern coast of the Huon Gulf. These rocks are also found in the north of New Caledonia, along the northeastern coast, with a direction N.20°–55°E.

If we leave New Guinea and pass to Australia, we see there the large central upland mentioned by Suess, which is composed of granite, gneiss, old schists, and a very extensive mantle of the "Desert Sandstone" (pp. 243–44). It extends to the western coast of Australia, where it is abruptly interrupted. In latitude 25°15′S. on this coast a Carboniferous series has been found (p. 243).

The eastern and southeastern parts of Australia, on the other hand, consist mostly of folded mountain ranges; there we find the Silurian, which is represented in Cape York peninsula (longitude $\pm 137\frac{1}{2}$°E., and latitude 35°S.), separating the gulfs of Spencer and St. Vincent (p. 249). North of this latter gulf, in the Flinders Range, there are outcrops of old schists and quartzites (p. 247). Further east, the old rocks of the Barrier and Grey ranges seem to be identical (p. 247); finally comes the great Australian Cordilleras, about which Suess writes (p. 248): "Des granites, des porphyres, des terrains cristallins, siluriens et dévoniens extrêmement plissés, généralement même redressés jusqu'à la verticale, constituent le noyau de la Cordillère. ... Le Carbonifère est presque horizontal."

I draw attention to the fact that in these mountain ranges the Silurian and the Devonian, although very much folded, are still recognizable by their fossils.

Let us cursorily follow Suess in his further views on the Cordilleras. With regard to Tasmania, he writes (pp. 251–52): "Du granite, des schistes cristallins et du Silurien s'y montrent redressés presque verticalement, suivant une direction méridienne; le Carbonifère, tant sous le faciès maritime que sous le faciès continental, recouvre en discordance ces terrains anciens." In the

[1] Suess, *op. cit.*, Vol. III, Part 3, pp. 1026–30.

eastern part of Victoria one meets again granite rocks and more or less vertical formations of gneiss and strata of Silurian age (p. 252). Continuing northward, we come to New South Wales, where north-south zones of granite and strongly folded Silurian and locally Devonian beds, together with old porphyry, form the eastern coast up to the Bateman Bay (latitude 35°40′S.). Northward this chain departs from the coast and disappears gradually under the horizontal Carboniferous (pp. 253–54). Further east another range consisting of granite and Silurian strata extends toward the north to Queensland (p. 254). In the northern part of New South Wales, the New England Range is composed of folded strata of Lower Paleozoic formations, with a nucleus of granite forming mountains (p. 254). To the north the granite diminishes, but the folded Devonian series increases (p. 255). To the west the Cordilleras disappears under the mantle of "desert sandstone" stretching to the Gulf of Carpentaria; at some points there rise through this mantle formations of granite and Paleozoic sediments (p. 255). Between latitude 22°30′ and 22°S. there are further patches of the Devonian and Silurian; but to the north of latitude 22°S. the coast, as far north as it is known, is formed of granite rocks, as also are the islands of the Strait of Torres, between Cape York and New Guinea (p. 256). In southeastern Australia the Silurian and Devonian, although much folded, have remained recognizable by their fossils up to latitude 22°S.

If we now pass to the opposite (northwest) side of the archipelago, to Southeastern Asia, what do we find? The islands of Banka and Billiton form a connection between the Dutch East Indies and the Malacca Peninsula. The granitic lands of these stanniferous islands extend in a northwestern direction over the southern point of the peninsula, in which (at Singapore) sandstones and old schists[1] have also been discovered. Just west of longitude 100°E. this granite chain turns to the north and ends in this direction at the island Kaw Tau, in the Gulf of Siam (Suess, Vol. III, Part 1, p. 303). To the west follows a depression in which flows the Bandon (p. 303). To the west of this depression comes another

[1] Suess, Vol. I, p. 600 and Vol. III, Part 1, p. 303. See also Verbeek, *Rapport sur les Moluques*, pp. 755–56, concerning the presence of Jurassic rocks in this series.

granite chain extending toward the north, thus forming a transition to the Birman chains (p. 303). At Tenasserim, on the neck of land east of the Mergui Islands, there are outcrops of Upper Carboniferous (p. 303) as well as of Archean rocks (I, 599). This Upper Carboniferous seems to extend far to the north in the valley of the Salwen, and at Moulmein notably (latitude $= 16\frac{1}{2}°$N.), in the form of limestone, it occupies a large area (p. 599). However, this eastern zone of the Birman chains, to which we have come, consists principally of Archean rocks (p. 593). To the east of Moulmein the chains extend northward and north-northeast, beyond Xieng-Sen, which is situated on the Mekong, in the northern part of Siam (Vol. III, Part 1, p. 290). Other ranges joining the former between Xieng-sen and Vien-tiane (also on the Mekong, but further down) take a northeast-southwest direction, and, decreasing in height, gradually disappear toward the southeast and the south, under the Siamese plain (p. 290). Then toward the southeast comes Siam, an immense unknown land, which, being connected with a vast area of granulite and granite, leads us to the low and old massif of Cambodge, of which we will speak later (pp. 290–91). Let us first follow the Birman chains toward the north.

According to Griesbach, the limestones of Moulmein seem to continue northward up to the highland of the Shan, where the Permian limestone is folded, the folds having been leveled (Vol. III, Part 1, p. 281). Then long ranges composed at first of crystalline schists and further north of gneiss lead north to Mandalay on the Irawaddi. And here again, in the Northern Shan States, we find once more the oldest fossiliferous strata which we had left in latitude 22°S. on the eastern coast of Australia. In fact, Suess writes (Vol. III, Part 1, p. 278): "Le chemin de fer et la route qui, de Mandalay, ont été construits vers le N.E. par Thibaw jusqu'au bac de Kunlon[1] sur la Salouen, longent sur de grandes distances la direction (N.N.E. ou E.N.E.) des roches. Les terrains les plus anciens sont à l'Ouest; au voisinage de Mandalay se montrent les couches inférieures de la série, peut-être cambriennes, puis le Silurien inférieur, le Silurien supérieur, un grès rouge pincé dans les plis,

[1] Kun-long, just to the south of the Tropic of Cancer.

et enfin la plaine récente du Salouen. ... *Le Silurien, affecté de plis, présente une grande variété de faciès.*"[1] The Silurian fossils of the Northern Shan States have not yet been found in the above-mentioned limestone series of the southern highland of the Shan (Vol. III, Part 1, p. 281).

From Kun-long, crossing the high ranges in a northwest direction up to Bhamo on the Irawaddi, we reach the mountains of the Siang-chan, which advance toward the north (p. 277). They consist of gneiss and crystalline schists and, for a great part, of granite (p. 277). On the west these mountains border on a zone of Carboniferous strata (p. 277). Then, turning again to the northeast, in the direction of Ta-li-fou, we recross the above-mentioned mountains of old crystalline rocks, which extend nearly to the Salwen in latitude about 25°N. Between this river and Young-tchang-fou, Lóczy found at Poupiao (latitude about 25°N.) the Silurian beds, whereas more to the east (eastward of Young-tchang-fou) he discovered Upper Carboniferous strata (pp. 276–77). This folded Paleozoic zone is situated between the Salwen and the Mekong (p. 276). East of the Mekong a broad zone of sandstone presenting the facies of the Flysch leads to the range of the Tsang-chan, consisting of crystalline schists. Next follows another sedimentary zone, composed principally of Upper Carboniferous strata which extend to the north, from Tali-fou to Batang (p. 276).

We need not continue in details our travel to the north and east, but we may limit ourselves to a few remarks. In the north the Upper Carboniferous strata play a somewhat important part, stretching themselves unconformably over older folded and denuded series of the Lower Paleozoic. After their deposition, there followed a new folding and, with regard to its tectonic history, this territory offers a fair similarity to many other countries of Eurasia (p. 276). Very far in the northeast of Central China, appear on one side the gneiss and crystalline schists, and on the other the oldest Paleozoic beds, all strongly pinched in successive mountain ranges. In the west, Silurian and Devonian series predominate (pp. 266–75); in the east, next to these, there are Cambrian strata (pp. 294–95). In Central and Northern China these formations of

[1] The italics are mine.

the Old Paleozoic attain a considerable development, as has been pointed out by von Richthofen,[1] Willis,[2] and others, as well as myself.[3] Concerning the Cambrian highland of Ordos, north of the Tsing-ling-chan, we refer also to Suess (Vol. II, pp. 303 ff., and Vol. III, Part 1, pp. 252 and 263). In that part of China unfolded Cambrian beds lie unconformably on folded and denuded Archean formations.

We need not go farther north, but now we return to Southern China, about which Suess (Vol. III, Part 1, p. 297) informs us: "Leclère[4] signale ... près de Hoaï-yuen, au Sud-Ouest de Kouen-lin, une arête de terrains précambriens et un culot de granite, accompagné, à l'Ouest, sur le Sikiang, par des roches précambriennes." And always following Suess, we proceed back again from Ta-li-fou to Cambodge, this time by the way of Yun Nan, Tonkin, and Annam. According to him (Vol. III, Part 1, p. 287), the mountain chains divide. The western ones, the Birman chains, we have followed in a zigzag way from south to north. The eastern chains lead in a southeastern direction to the Cordilleras of Annam. Without entering into details, we may say that, between Ta-li-fou and Indo-China, the oldest fossiliferous rocks continue gradually to disappear. It is only more eastward, at Lou-nan (east of Yun-nan-fou), that Leclère recognized series of the Middle and the Upper Devonian, together with various formations of the Carboniferous and the Permian (p. 296). The tectonic relation between these rocks and those of the above-mentioned pre-Cambrian (also discovered by Leclère) cannot yet be established. We read in Suess (Vol. III, Part 1, p. 297): "Les relations de ces anciens terrains ne se dégagent pas encore avec netteté des documents publiés jusqu'à présent. Il n'est guère possible de dire

[1] F. von Richthofen, *China*, 5 volumes and one atlas; Vol. I, 1877; Vol. II, 1882; Vol. IV, 1883, atlas North China, 1885; Vols. III, V; and atlas, South China, 1911.

[2] B. Willis and E. Blackwelder, *Research in China*, 2 volumes and one atlas; Vol. I, 1907.

[3] E. C. Abendanon, "La Géologie du Bassin Rouge de la province du Se-Tchouan," *Rev. Univ. des Mines*, etc., Liège, 1906. E. C. Abendanon, "Structural Geology of the Middle Yang-tzikiang Gorges," *Journal of Geology*, 1908.

[4] A. Leclère, "Etude géologique des provinces chinoises voisines du Tonkin," *Annales des Mines*, 1900–1901.

s'ils auprésentent le substratum de la plate-forme calcaire comme auprès d'I-tchang[1] ou s'ils font partie du massif ancien qui existe probablement dans la Sud-Est de la Chine, en formant un avant-pays indépendant des Altaïdes."

The most recent data about a fairly considerable area of the eastern part of the Chinese province of Yun Nan are supplied by Deprat.[2] This explorer gives many details. He writes (p. 44):

> Les terrains métamorphiques ou granitiques n'apparaissent que dans le S., dans la vallée du Fleuve Rouge. ...
>
> Le Cambrien offre un développement énorme au Yun-Nan, aussi bien dans son extension verticale qu'horizontale; *les séries fossilifères sont abondantes.*[3]
>
> J'ai découvert l'Ordovicien au N. d'Yi-léang. Le Gothlandien paraît représenté dans la même région.
>
> Le Dévonien complet offre un développement colossal au Yun-Nan. J'ai reconnu le Dévonien inférieur. Dans le Mésodovénien, l'Eifélien et le Givétien sont bien caractérisés, de même que le Frasnien et le Famennien dans le Dévonien supérieur.
>
> Nous avons pu affirmer la présence du Dinantien, représenté par plusieurs niveaux. Le Moscovien offre un développement énorme, ainsi que l'Ouralien.

About the relation between all these formations, Deprat goes into details, too, and he informs us that the structure of this territory is very complicated. According to him (pp. 251–60), a local orogenic movement (in the southeastern part of Yun Nan) took place after the Cambrian; there was uninterrupted sedimentation during the Silurian, and the corresponding beds ought to exist from Kwei-chow in the east to Burma in the west. The sedimentation continued until the Muscovian in well-developed series. Then there followed a strong ante-Uralian folding; the Uralian formation lies unconformably on the denuded folds of the Old Paleozoic. At the end of the Permian, when marine sedimentation came to an end and before any deposit of Lower Triassic sediments had taken place, a very intense orogenic movement occurred. After that, Deprat writes (p. 252): "La dernière

[1] Situated on the Yang-tzi-kiang, at the point separating the middle from the lower part of this river.

[2] J. Deprat et H. Mansuy, *Etude géologique du Yun-Nan oriental*, Vol. I, Fasc. I, 1re partie, Hanoï-Haiphong, 1912.

[3] The italics are mine.

phase principale de plissement, dont l'intensité a été véritablement très grande, est la phase de dislocation post-triasique, dont les effets ont été considérables; je la considère avec M. Lantenois comme une phase himalayenne: (p. 257) l'ensemble du Yun-Nan est alors plissé d'une façon très énergique, les charriages y prennent une amplitude très grande et la région du haut Fleuve Bleu (Yang-tsé-Kiang), prolongement sud du Yung-ling-chan, est charriée sur la région yunnannaise." The last movements were due to the upheaval which raised Yun Nan to the relatively high altitude which it occupies nowadays. They occurred during the Plio-Pleistocene.[1]

It is clear that the tectonic movements of Yun Nan, where the excessive thick Old Paleozoic series appear on such an extensive area and with a particularly great richness of fossils, have been extremely intense and numerous.

In Northern Annam, F. Deprat[2] discovered at Ben-Thuy the Ordovician, and he believes that this formation will also be discovered in Tonkin and Laos. The Devonian and Dinantian are also represented in this region, and the Uralian is once more in transgression (exactly as in Yun Nan) upon a folded and denuded territory of the older Paleozoic.

In the same way as in the Northern Shan States far to the west, here in the east of Southeastern Asia we find the most southern country where fossils of the Old Paleozoic have been discovered; for according to Suess, in the plain of Tonkin soft schists have been assigned to be Devonian; but this point is far from being sure (Vol. II, p. 277). In the island of Haï-Nan, we find Suess mentioning (Vol. III, Part 1, p. 297) that according to Madrolle, the center of the island is composed of granite and schists, and is

[1] These last movements of epirogenic character, giving rise to the phenomena of what I have called anticlinal distraction (in a booklet titled *Die Grossfalten der Erdrinde*, Leyde, Holland, 1914), which phenomena led to the forming of important tectonic depressions, are so completely similar to those of the Plio-Pleistocene period in Central Celebes, where I have classified them together in a single process of grand-folding, that I do not hesitate to identify the most recent epirogenic movements of Yun Nan with those of Central Celebes, the latter having induced the upheaval of this country to its present high altitude.

[2] "Sur la découverte de l'Ordovicien, à Trinucleus, et du Dinantien dans le Nord-Annam, etc.," *Comp. Rend. Ac. d. Sc.*, 28–5–12, T. 154, pp. 1452–54.

surrounded by a sandy argillaceous formation of red color in the north and yellow in the east.

Patches of Carboniferous limestone occur more abundantly as we approach the Cordilleras of Annam, on the east coast. Suess mentions (Vol. III, Part 1, p. 297) that Bell crossed these Cordilleras in a westerly direction, starting from Tourane (on the eastern coast of Annam, by latitude 16°10′N.). They appear to be composed of various zones, first of dioritic and granitic rocks, then of gneiss and crystalline schists, and lastly of rather recent formations (Vol. II, p. 275, and Vol. III, Part 1, p. 297). The granitic rocks advance toward the south, along the coast, from latitude 14°N. to Cape St. Jacques, southeast of Saigon, and they must be the southern extremity of a very long mountain chain stretching in a northern direction to the granite highland of Laos (Vol. II, p. 275).

In this way we come back to the old low massif of Cambodge and to the delta of the Mekong, in Cochin-China.

Like Australia to the southeast, Cochin-China and Tonkin constitute very old uplands (Suess, Vol. I, p. 607) to the northwest of the East Indian Archipelago. Between these uplands we have the granite massifs of southwest Borneo, which separate the folds of the eastern part of this island into two zones, one extending toward the west and the other toward the south, the massifs of granite, gneiss, and crystalline schists of Central Celebes and the horsts of Boeroe and Ceram which, together with others, are probably part of an old massif (Suess, Vol. III, Part 3, p. 1036).

It is only southward from latitude 22°S., along the eastern coast of Australia on one side, and northward from the Shan states, North Annam, and Yun Nan on the other side, that the Silurian and Devonian appear. Then, on the northern side, more to the north, exists the Cambrian and the pre-Cambrian beneath. The latter plays an important part to the south, but still more north of the Tsing-ling-chan in China. On the southern side (viz., in Australia), however, these last-mentioned two formations do not appear.

In the intermediate territory, to which the archipelago belongs, outcrops of old crystalline schists occur in many places, but no explorer has discovered any fossils of the Old Paleozoic.

The Old Paleozoic series which border the periphery of the above-mentioned territory are strongly folded. The geo-metamorphic processes in those areas, that are nevertheless characterized by their fossils, have certainly not been less powerful than in the intermediate territory stretching from latitude 20°N. to latitude 22°S., over the southeastern part of Asia, the Dutch East Indian Archipelago, New Guinea, and part of Australia.

Is it then credible that in this extensive territory, the extent of which from southwest to northeast is not known, all fossils of the Old Paleozoic should have disappeared? What particular reasons not valid for Southeastern Asia and Eastern and Southeastern Australia, could plausibly explain their disappearance? Is it not more logical to consider, under these circumstances, the absence of fossils, not as an indirect proof, but as a powerful argument for the hypothesis that in the above-outlined territory there never existed any Old Paleozoic fossils?

Is it not in the same way remarkable that the same transgression of the Upper Carboniferous and the Permain over the folded and denuded folds of the Old Paleozoic, which has been observed as well in Southeastern Asia as in Eastern Australia, is also apparent in some parts of the intermediate territory (for instance, in the west of Sumatra and in the island of Timor), whereas fossils of the Old Paleozoic have nevertheless not been found there?

From the above-stated considerations it seems not only possible to deduce a precise conclusion about the age of the crystalline schists formation, but also to deduce others of essential interest. They are the following:

1. The gneiss, the mica schists, the phyllites, and the real "old" schists (thus with the omission of the rocks which do not make part of them) must be Archean and pre-Cambrian rocks.

2. They once built up an Old Paleozoic continent, which extended at least over an area of 45° in latitude, between the tropics, from the southeast of Asia to the east of Australia. Its development from southwest to northeast is unknown, owing to the presence of the Indian and Pacific oceans, but at all events this continent must have included most of Sumatra on the southeast, and the Philippine Islands on the northeast, considering that, in

those countries too, there has not yet been found any fossil of the Old Paleozoic. To the west, it may have stretched out as far as Madagascar.

3. In the central part, north and south of the equator, mountain ranges of an almost east-west direction must have played an important part in this very old continent.

4. The Cambrian deposits on the north leaned against it, but in the south they have not been found. Nevertheless they may have existed south of Australia, where the ocean now is.

5. In the north, to a large extent toward the south, and in the south, fairly far toward the north, a transgression passed over the border areas of this continent during the Silurian, the Devonian, and the Lower Carboniferous periods (with a secondary unconformability).

6. Then a folding occurred of these sediments of the Old Paleozoic. Border ranges were formed at the periphery of the old continent which, from a tectonic point of view, are similar to the Tertiary Cordilleras.

7. During the Middle Carboniferous, denudation and leveling took place.

8. Lastly, the transgressions of the Upper Carboniferous and the Lower Permian seas invaded (with a real unconformability), not only the denuded Eopaleozoic border ranges, but also part of the old continent itself.

To this continent, the existence of which during the Old Paleozoic seems to be evident, I give the name of *Aequinoctia*.

It was during the Permo-Carboniferous that its dislocation began, in the territory of the Dutch East Indian Archipelago. The demembration of this oldest continent must have continued during the Mesozoic and the Tertiary; and so it became, in time, one of the most unsteady parts of the earth's crust. In this connection it is interesting to observe that the territory of the archipelago exhibits nowhere the tremendous development of the Mesozoic which has been observed, for instance, in the Alps, though uninterrupted sedimentation has taken place in some regions during the whole Mesozoic and part of the Tertiary. As a rule, however, complete series appear hardly anywhere. This

peculiarity can best be explained, I venture to suggest, by the aforesaid demembration of the original old territory.

A last question before concluding: Is it probable, in the future, that there may still be discovered in this territory fossils of the Old Paleozoic? Will it be the same with *Aequinoctia* as with the Jurassic Sino-Australian continent of Neumayr?

The numerous explorers who have worked in so many different parts of this region, still so unknown in the time of Neumayr, have succeeded in the course of years in making known, as we have seen, in a fairly continuous series, every period from the Upper Carboniferous to the Recent. Older fossils have not as yet been discovered, notwithstanding the greatest efforts and the closest attention, and in spite of the fact that the rocks preceding the Upper Carboniferous outcrop in large areas and are frequently deeply cut by erosion, thus offering very favorable points of investigation. Under these circumstances, I suppose that the hypothesis of the existence of *Aequinoctia* during the Old Paleozoic may be maintained.

ON COARSE GABBROID DIABASE IN WESTFIELD, MASSACHUSETTS[1]

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In the western part of the town of Westfield, Massachusetts, a number of quarries have been opened in the Holyoke diabase sheet. These quarries are the same which have furnished the beautiful datolite specimens which are to be found in all collections. The diabase or trap in which these quarries are located is the middle or main extrusive trap sheet of the Connecticut Valley Triassic area. Here the main portion of the sheet is composed of a gray holocrystalline diabase of fine to medium grain not unlike the intrusive traps which form East and West Rocks near New Haven. In the northernmost of the four quarries there occur, included in the normal rock, irregular areas of a gabbroid rock very unusual in appearance for an extrusive lava. While studying the occurrence of datolite and other secondary minerals in these quarries, the writer's attention was attracted to these very coarse-grained phases of the sheet and they were examined in some detail. At first it was believed that they represented a later intrusion which had chanced to penetrate the previously existing flow, but a study of the relations of the coarse material to the surrounding diabase of normal grain led to the abandonment of this view. The contact with the surrounding rock is sharp, the transition from coarse to fine material being accomplished in a distance of an inch or less. The coarse-grained gabbro forms irregular rounded areas often many yards in diameter in the finer-grained diabase. In the hand specimen this coarse rock shows broad blades of bronzy greenish-black pyroxene which reach an inch or more in length, imbedded in a coarse granular aggregate of pyroxene and greenish feldspar. Under the microscope the rock is seen to be composed of large

[1] Published by permission of the Secretary of the Smithsonian Institution.

crystals of pyroxene and plagioclase in a subordinate fine-grained ground-mass composed largely of needle-like crystals of plagioclase with interstitial augite in anhedral patches now largely altered to a greenish-brown chlorite which is probably diabantite.[1] The large pyroxene phenocrysts are rather pale greenish-brown in color, with well-developed cleavage. They are unaltered and are in part transparent and in part turbid from the presence of an opaque brownish pigment. They are not commonly simple but are bladed aggregates made up of a number of irregular interlocking crystals not quite in parallel position so that they do not all extinguish simultaneously. In composition this pyroxene is ordinary augite. The plagioclase phenocrysts are characterized by rather narrow twinning lamellae and are near-basic andesine in composition. The most striking microscopic feature of the rock is the presence of broad patches of a graphic intergrowth of plagioclase and augite in fixed orientation, giving remarkable pegmatitic textures entirely analogous to the quartz-feldspar aggregates of graphic granite. In places a phenocryst of plagioclase terminates in a branching fernlike intergrowth of plagioclase and augite.

A question naturally arises as to the origin of such coarse forms in an intrusive flow less than a hundred feet in thickness. Ordinarily such surface flows cool too rapidly to permit the growth of large crystal individuals. The mode of occurrence of this coarse phase indicates that the large crystals of plagioclase and augite were formed after the lava had reached its present position and precludes the supposition that it was intruded into the extrusive sheet after the sheet had been buried by later sedimentation. Emerson[2] has described similar coarse forms from this trap sheet further north associated with mudstone, pitchstone, etc., formed by the molten lava coming in contact with water and mud. Strangely enough, he attributes the coarseness of grain of the plumose diabase to sudden cooling or quenching of the molten

[1] Earl V. Shannon, "Diabantite, Stilpnomelane, and Chalcodite from Westfield, Massachusetts," *Proc. U.S. Nat. Mus.* In print.

[2] B. K. Emerson, "Plumose Diabase and Palognite from the Holyoke Trap Sheet," *Bull. Geol. Soc. Amer.*, XVI (1905), 91.

material by the water. This is quite contrary to our knowledge of the effect of the rate of cooling upon the grain of igneous rocks. The writer after some study has reached the conclusion that the coarseness of crystallization of the gabbroid diabase is due to absorption of confined water by the still molten portion of the flow, with the formation of miniature pegmatite chambers. Quite certainly here as elsewhere the lava of this flow encountered moist mud flats and ponds of water. The coarse gabbro occurs for the most part near but not at the top of the flow. If we may conceive the top of the flow as forming here a resistant crust due to early solidification, the water from below heated to a high temperature and ascending through the still molten lava must come to rest beneath the previously solidified crust and might readily be conceived to mix with the still molten material to form a magma rich in volatile constituents and thus capable of remaining fluid for longer periods and at lower temperatures than when in its original dry state. This would permit the formation of coarse-grained pegmatitic textures like those here described. Emerson has shown in the paper cited that water did ascend through the flow and that such water in some instances caused explosions rupturing the upper crust. It is easy to understand how pressures could obtain in a pegmatitic chamber of the sort indicated sufficient to rupture the roof of solid diabase after the formation of large plumose crystals of augite was well advanced. The remaining molten material would be chilled following the explosion and yield perlitic glasses associated with plumose diabases exactly in the manner described by Emerson.

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